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Tomus XXXV.

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NOTE TO CONTRIBUTORS

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Methods, techniques, material studied, description of the area investigated, etc.
Results
Discussion or conclusions
Acknowledgement
Explanation of plates (if any)
Tables
Captions of figures (drawings, photomicrograps, etc.)

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ARSENATE MINERALS FROM HUNGARY

S. SZAKÁLL*

Department of Mineralogy, Herman Ottó Múzeum

W. D. BIRCH**

Department of Mineralogy, Museum of Victoria

Á. KOVÁCS***

Department of Metallurgy, University of Miskolc

W. POSTL****

Department of Mineralogy, Steiermärkisches Landesmuseum Joanneum

ABSTRACT

In this paper we introduce some arsenate minerals found recently in the ore deposits and ore indications in Hungary. The essential demand of their formation was the presence of arsenopyrite or the members of the tetrahedrite group. Under such circumstances it was generally possible to identify arsenate minerals from the oxidized zone of the ore deposits. Arseniosiderie, segnitite, scorodite and probably arsenbrackebuschite were found in the hydrothermal polimetallic ore veins at Nagybörzsöny, Börzsöny Mts. Olivenite, mimetite, beudantite and a tyrolite-like arsenate mineral were found in the hydrothermal-metasomatic iron ore deposit with accessory at Rudabánya, Rudabánya Mts. In a limestone quarry near Pécs, Mecsek Mts. olivenite, conichalcite and tyrolite were found. Talmessite and picropharmacolite were identified in dolomite concretions in the sedimentary infiltrated uranium ore deposit at Kővágószőlős, Mecsek Mts. Scorodite and pharmacosiderite were observed in the epithermal alunite veins of the limnoquartzite at Mád, Tokaj Mts. Pharmacosiderite was identified in the epithermal mineral assemblage of hydroquartzite at Gyöngyössolymos, Mátra Mts.

INTRODUCTION

Arsenate minerals have recently ben discovered in Hungary. Annabargite from the serpentinite at Helesfa (Mecsek Mts.) was referred to by SZEDERKÉNYI (1962).

According to the literature (KOCH 1985, PAPP and WEISZBURG 1986) the members of the tetrahedrite group are widely distributed, in small quantity though. A lot of microprobe analyses were made on them (DOBOSI 1976, 1979, 19880, 1982, 1983, 1984) which showed that the As content could reach up to 20 percent in some cases.

* H–3525 Miskolc, Kossuth u. 13.
** Melbourne, Victoria 3000. Russel Street 285, Australia
*** H–3515, Miskolc, egyetemváros
**** A–8010 Graz, Raubergasse 10, Austria

It could be expected that in the oxidized zone of the ore deposits arsenate minerals would appear. The same is the situation at the Nagybörzsöny ore veins being the only place in Hungary where arsenopyrite is found. It is long been known that arsenopyrite is one of the dominant sulphide in the ore veins at Nagybörzsöny (KOCH and GRASELLY 1952, PANTÓ and MIKÓ 1964).

The detailed description shows that the arsenates have the same type of occurrence – they are found in the oxidation zone of the deposits containing arsenopyrite and/or the members of the tetrahedrite group (Table 1. *Fig. 1*)

TABLE 1

Arsenate minerals from Hungary

LOCALITIES / SPECIES	Nagy-börzsöny	Ruda-bánya,	Pécs	Kővágó-szőlős	Mád	Gyöngyös-solymos	Helesfa
Annabergite							+
Arsenbrackebuschite ?	+						
Arseniosiderite	+						
Beudantite		+					
Conichalcite			+				
Mimetite		+					
Olivenite		+	+				
Pharmacosiderite					+	+	
Picropharmacolite				+			
Scorodite	+				+		
Segnitite	+						
Talmessite				+			
Tyrolite			+				

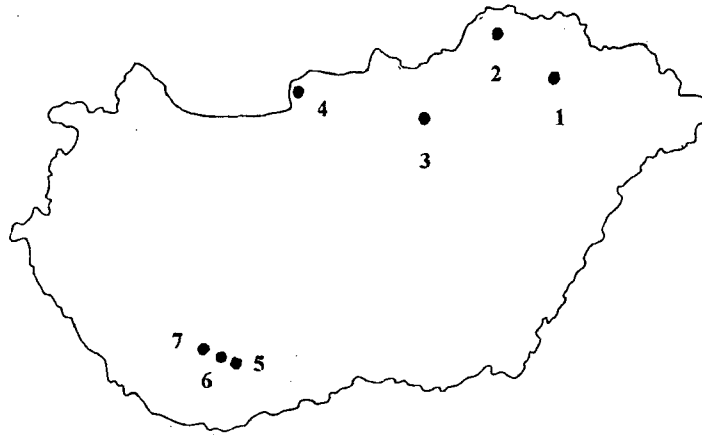

Fig. 1. Occurrences of arsenate minerals from Hungary
1. Mád 2. Rudabánya 3. Gyöngyössolymos 4. Nagybörzsöny 5. Pécs 6. Kővágószőlős 7. Helesfa

EXAMINATION METHODS

X-ray powder diffraction test: SIEMENS D500 diffractometer, Si sample container, Cu K_α radiation. 35 kV accelerating voltage, 30 mA tube current, 0,05–1,00 gard/sec goniometer speed. Diffrac. computer aided elaborating system. The tests were made at the Dept. of Mineralogy of the Steiermärkisches Landesmuseum Joanneum, Graz, Austria. (Some of the test were made elsewhere, this are named at the place of reference).

Quantitative microprobe analyses: CAMECA CAMEBAX SX50. Tests circumstances: 25 kV accelerating voltage, 0,1 mA test current. Standard materials: wollastonite (Si, Ca), jadeite (Na), synthetic $KTaO_3$ (K), hematite (Fe), corundum (Al) benitoite (Ba), galena (Pb), sphalerite (Zn), metallic copper (Cu), metallic manganese (Mn), fluorapatite (P), arsenopyrite (As). The test were made at the Dept. of Mineralogy of the Museum of Victoria, Melbourne, Australia (measurements in order to determine H_2O content were not made).

Scanning electron microscope examinations, qualitative microprobe analyses: AMRAY 1830 I., connected EDAX 9900 energy dispersive microprobe analyser. Test circumstances: 20 kV accelerating voltage, 10^{-10}. A test current, SiLi detector, incline angle 35 DEG, W cathode, EDAX-EDS system. Examinations were made at the Dept. of Metallurgy, University of Miskolc, Hungary.

RESULTS

NAGYBÖRZSÖNY

The mineralogical description of the hydrothermal ore veins at Nagybörzsöny was based upon the studies of KOCH and GRASSELLY (1952) and PANTÓ and MIKÓ (1964). These papares revealed that one of the dominant minerals of the ore deposit is arsenopyrite. The minerals of the tetrahedrite group are much less distributed. Therefore the formation of arsenates is generally to be connected to the occurrence of arsenopyrite. We found samples on the waste dumps at Altáró, Afelső-Rózsa and Alsó-Rózsa adits which show that the primary sulphides dissolved, and the secondasy sulphates oxides and smaller quantity of arsenates were formed. The cavities of the quartz veins contain scorodite, arseniosiderite, segnitite and probably arsenbarckebushite accompanied by quartz, gypsum, goethite and anglesite. Right next to the cavities arsenopyrite, pyrite and galena can be observed.

Arseniosiderite

Forms rounded masses crust built up from brown crystals smaller than 1 milimeter on the waster dump of Altáró. Its accessory minerals are rhombohedral segnitite and lamellar crystals of a mineral thought to be arsenbrackebuschite. Their structure can clearly be seen on the SEM micrograph (*Fig. 2*). Its powder diffraction examination (Table 2) shows that the strongest reflections distinctly appear, the intensity ratios are somewhat different compared to those of the correspondant JCPDS card. Its quantitative microprobe analysis (Table 3) revealed that the arseniosiderite from Nagybörzsöny contains more As and less Fe than theat introduced in the study of DANA (1920).

TABLE 2

X-ray powder diffraction data of arseniosiderite from Nagybörzsöny

arseniosiderite Nagybörzsöny		arseniosiderite JCPDS 13-134	
d(Å) (obs.)	I/I_0	d(Å)	I/I_0
8.93	100	8.95	100
2.96	67	2.95	80
2.78	42	2.77	100
2.50	19	2.50	90
2.22	32	2.20	70
1.355	12	1.350	70

Made in Steiermärkisches Landesmuseum Joanneum, Graz

TABLE 3

Chemical composition of arseniosiderite from Nagybörzsöny (Wt%)

CaO	15.04
Fe_2O_3	31.90
PbO	0.30
P_2O_5	0.23
As_2O_5	45.50
Total	92.97

Analyst: BIRCH, W. D. (Museum of Victoria, Melbourne)

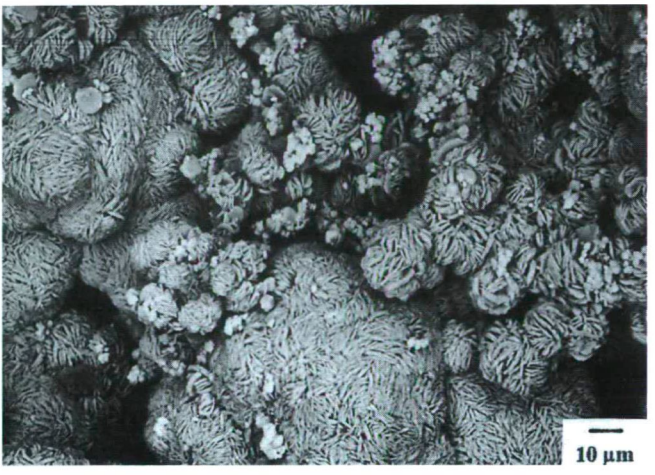

Fig. 2. Arseniosiderite, globular crystal groups, Nagybörzsöny. Scanning electron micrograph

Segnitite

It is found in the same environment as brownish-yellow crusts and sometimes forms 0,X mm rhombohedrals. The SEM micrograph shows that the crystals are built up from (1011) planes, the (0001) plane appears as well (*Fig. 3*). Segnitite was still referred to as baudantite in the study of SZAKÁLL (1992), since there was not quantitative chemical analysis made on the sample that time. Its powder diffractions record (BIRCH *et al.* 1992) (Table 4) gave no clear evidence for the identification. Eventually its quantitative microprobe analysis (Table 5) clarified that the sample was segnitite, as hardly any (0,55% and 2,36%) SO_3 was found in it.

TABLE 4

X-ray powder diffraction data of segnitite from Nagybörzsöny

segnitite Nagybörzsöny		segnitite BIRCH *et al.* (1992)		beudantite JCPDS 19-689	
d (Å) (obs)	I/I_0	d(Å)	I/I_0	d(Å)	I/I_0
5.99	64	5.966	50	5.99	80
5.69	13	5.719	10	5.72	20
				5.13	5
3.68	44	3.678	40	3.67	70
3.54	13	3.530	5	3.54	20
		3.119	5		
3.09	100	3.092	100	3.08	100
		3.011	10		
2.96	15	2.987	5	2.97	30
2.84	24	2.849	20	2.84	50
2.55	18	2.550	10	2.54	50
2.38	9	2.382	5	2.37	20
		2.334	2		
2.31	9	2.318	3	2.31	20
2.27	25	2.283	30	2.27	60
2.25	12	2.254	15	2.24	15
				2.12	10
		2.011	5	2.09	10
1.990	20	1.992	30	1.979	60
1.840	14	1.840	25	1.829	60
1.772	6			1.768	10
				1.745	5
				1.706	20
				1.694	10
1.687	10	1.688	15		
				1.678	40
				1.641	10
				1.560	20
1.545	10	1.546	8	1.536	40
1.503	7	1.508	10	1.499	40

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TABLE 5

Chemical compositon of segnitite from Nagybörzsöny (Wt%)

CuO	0.10	– 0.09
BaO		0.09
SrO	–	0.08
PbO	28.45	28.27
Fe_2O_3	31.62	35.17
Al_2O_3	0.19	0.15
P_2O_5	0.41	0.16
As_2O_5	26.17	29.01
SO_3	2.36	0.55
Total:	89.48	93.48

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

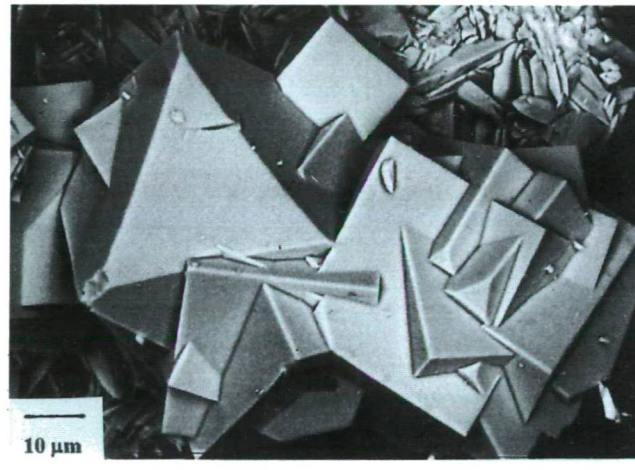

Fig. 3. Segnitite, rhombohedral crystals. Nagybörzsöny. Scanning electron micrograph

Arsenbrackebushite (?)

Accompanying segnitite and arsenbrackebuschite, 10–15 μm large lamellar aggregates with Pb, Fe content were observed (*Fig. 4*). Analysing the arsenates samples we found phases with a chemical composition that best correspond with arsenbrackebuschite (Table 6). The essential difference comes from the smaller Zn and higher Fe content in them. Its powder diffraction tests have not been made yet, because of the very smal quantity of the samples, so its presence is not proved. ABRAHAM *et al.* (1978) described this mineral first from a similar paragenesis with anglesite and beudantite.

TABLE 6

Chemical composition of arsenbrackebuschite (?) from Nagybörzsöny (Wt%)

CuO	–	–
ZnO	0.19	0.14
CaO	–	–
Fe_2O_3	12.78	16.26
Al_2O_3	0.45	0.21
P_2O_5	0.32	0.27
As_2O_5	29.25	30.53
SO_3	1.37	1.05
Total	97.56	97.08

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

Fig. 4. Arsenbrackebuschite (?), lamellar crystals with segnitite, Nagybörzsöny. Scanning electron micrograph.

Scorodite

It is most distributed arsenates mineral at Nagybörzsöny. Its found on the waste dunps of Altáró adit, Felső-Rózsa adit and Allsó-Rózsa adit. It occurs in diverse appearance in the cavities of quartz veins. It forms green, yellowish-green crusts, rounded masses and several mm thick earthy covering. Sometimes it also forms banded structures. On the waste dump of Alsó-Rózsa adit it also appears as light green earthy masses in the fissures of arsenopyrite. For it always accompanies arsenopyrite, its formation can be derived from the weathering of that mineral. According to the SEM examinations the surface of the rounded masses is made out from lamellar crystals (*Fig. 5*) The powder diffraction data correspond well with that of its JCPDS card, so it is not needed to present them in this paper. Its microprobe analysis (Table 7) data nearly identical with those in the literature (DANA 1920) apart from somewhat more Fe.

TABLE 7

Chemical composition of scorodite from Nagybörzsöny (Wt%)

CaO	0.07
Fe_2O_3	38.43
Al_2O_3	0.19
P_2O_5	–
As_2O_5	51.65
Total	90.34

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

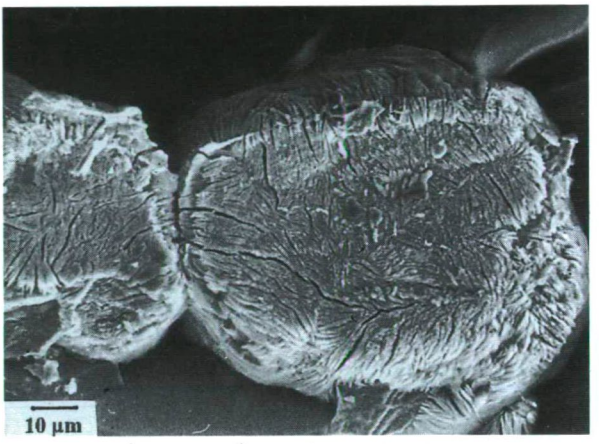

Fig. 5. Scorodite, lamellar crystal groups, Nagybörzsöny. Scanning electron micrograph.

RUDABÁNYA

The presence of arsenic as a trace element has long been well known in th primary and secondary zones of the hydrothermal metasomatic iron ore depostit with sulphide minerals (PANTÓ 1956). Based upon the microprobe analyses conducted by DOBOSI and NAGY (1982), found first an As containing mineral (Hg-tennantite) from a siderite body in the oxidation zone of the deposit. As containing minerals of secondary origin (olivenite and mimetite) from the oxidation zone were first described by SZAKÁLL (1992, 1994).

Beudantite

Its found as a member of a mineral association rich in species with Cu, Ag, Sb, S, Cl, Br, I elements in the silicious limonite orebodies in the Adolf area of the deposit. The rock primarily consists of quartz, and cerussite can often be found in it as xenomophous or isomorphous grains reaching 1 cm. Beudantite was observed in the cerussite as minute – 10 μm grains – (*Fig 6.*) and in 1–3 mm crystalline patches surrounded by

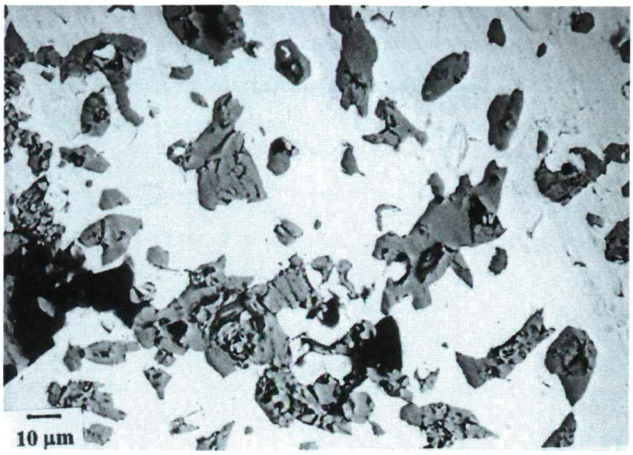

Fig. 6. Beudantite, inclusions in cerussite, Rudabánya, Adolf mine. Scanning electron micrograph.

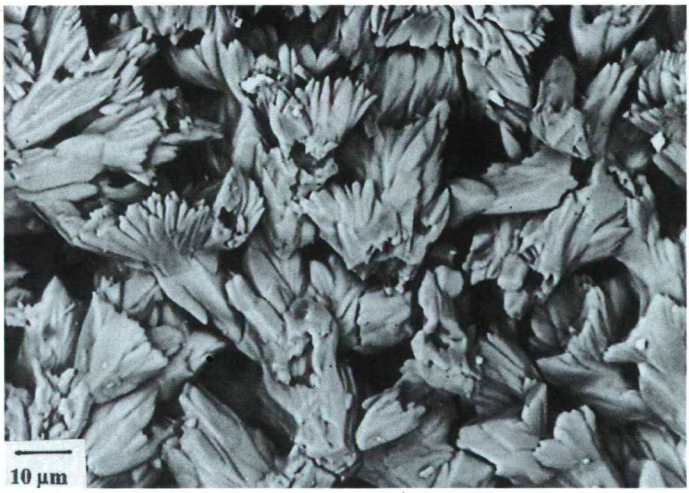

Fig. 7. Beudantite, finger-like aggregates, Rudabánya, Adolf mine. Scanning electron micrograph.

cerussite. The yellow finger-like aggregates make up 0,X mm crystals groups (*Fig. 7*). The 0,X mm large, short, columnar mimetite crystals can often be observed in the beudantite aggregates. Its X-ray examination show good correspondance not only with the reference card of beudantite but that of segnitite as well (Table 8). Having regard to be fact however that the quantitative microprobe analysis pointed out its nearly 10% SO_3 content (Table 9.), it can only be beudantite.

TABLE 8

X-ray powder diffraction data of beduantite and mimetite from Rudabánya

Beudantite Rudabánya		Beudantite JCPDS 19-689		Other minerals
d(Å) obs	I/I_0	d(Å)	I/I_0	
5.97	41	5.99	80	
5.69	2	5.72	20	
4.44	3			M
4.26	9			Q
3.67	43	3.67	70	
3.55	5	3.54	20	
3.34	44			M, Q
3.09	100	3.08	100	
3.05	24			M
3.01	30			M
2.98	19	2.97	30	
2.96	14			M
2.85	14	2.84	50	
2.55	14	2.55	50	
2.38	8	2.37	20	
2.31	8	2.31	20	
2.28	23	2.27	60	
2.25	8	2.24	30	
2.12	5	2.12	10	
2.11	6			M
2.09	4	2.09	10	
1.988	31	1.979	60	
1.965	8			M
1.936	4			M
1.907	9			M
1.837	27	1.829	60	
1.817	11			Q
1.713	5	1.706	20	
1.687	7	1.678	40	
1.574	4	1.560	20	M
1.543	12	1.536	40	
1.506	7	1.499	40	
1.416	3	1.409	20	
1.383	3	1.385	20	
1.349	8	1.344	40	

Made in ALUTERV-FKI (Budapest)
(M=mimetite, Q=quartz

TABLE 9

Chemical compostition of beudantite from Rudabánya (Wt%)

ZnO	0.32
PbO	30.71
Fe_2O_3	32.71
Al_2O_3	0.11
As_2O_3	13.47
SO_3	8.35
Total	85.68

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

Ca-Cu containing arsenate mineral

In the fissures and cavities of the altered siderite ore of the Andrássy III area of the deposit turquoise-blue spherical aggregates of a tyrolite-like mineral reaching 0,X mm were observed. According to the SEM images the aggregates are made up from tabular crystals (*Fig. 8*). Its X-ray test revealed that its structure is resemble to tyrolite, but there are a lot of difference and it cannot be matched to any arsenate known so far. Its detailed examination has not been finished yet.

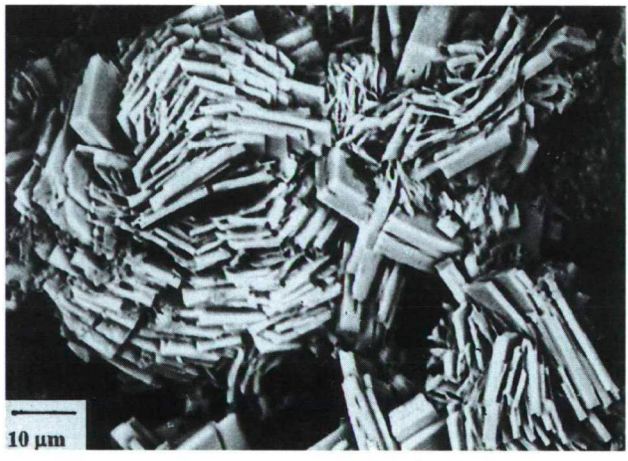

Fig. 8. Tyrolite-like mineral, Rudabánya, Andrássy III. mine. Scanning electron micrograph.

PÉCS

There is a copper indicating locality described by TOKODY (1952) in a limestone quarry near the Kozári hunting lodge. Small quantity of arsenate minerals has recently been observed with azurite and malachite. Sulphides producing secondary Cu-minerals were also found in the brecciated zones. These are chalcocite and tennantite idetified

with X-ray and microprobe tests. Both sulphides appear as small massive patches reaching some mm in the breccias and if often infiltrates into the limestone. The weathering processes of tennantite produced the observed arsenates: conichalcite, olivenite and tyrolite.

Conichalcite

It generally appears as yellowish-green crusts, mostly on the surface of calcite crystals. It was also observed in pseudomorphous columnar habit. We suppose it had encrusted azurite crystals then the azurite dissolved and left behind the conichalcite crust. It rarely can be found as tufted masses accompyying malachite and olivenite (*Fig. 9*). The four strongest reflections appeared on the X-ray diffraction record (Table 10). The results of the microprobe analysis (Table 11) show good correspondance with those of other tests (DANA 1920).

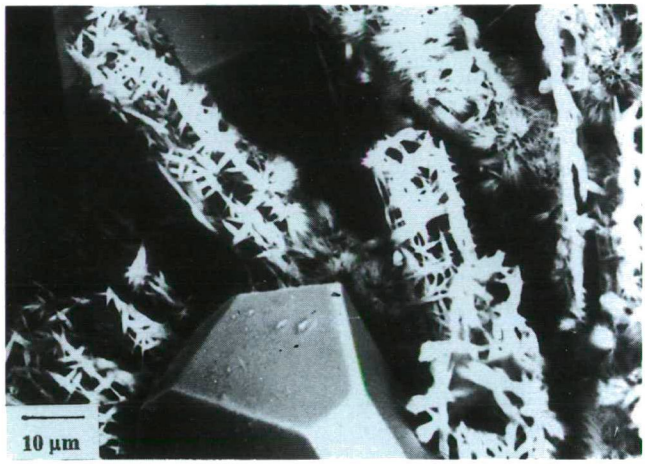

Fig. 9.
Conichalcite, sparys on needles olivenite crystals, Pécs, Kozári vadászház quarry.
Scanning electron micrograph.

TABLE 10

X-ray powder diffraction data of conichalcite from Pécs

Conichalcite Pécs		Conialcite JCPDS	
pos(DEG)	C I/I_1	Pos(DEG)	I/I_1
28.586	89.6	28.569	80
31.488	100.0	31.464	100
34.550	72.8	34.440	65
		34.660	55
35.156	64.8	35.108	45

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TABLE 11

Chemical composition of conichalcite from Pécs (Wt%)

CuO_2	31.14
ZnO	0.44
BaO	0.17
Al_2O_3	0.23
PbO	0.46
CaO	19.66
P_2O_5	0.48
As_2O_5	37.32
Total	89.90

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

Olivenite

It was rarely been observed like conichalcite. It forms light green columnar crystals which make up tightly fitting groups. The SEM image of olivenite is shown on *Fig. 10*, on which the (110), (011) planes can be observed. 23 peaks appeared on the X-ray diffraction record and the intesnity rations correspond well with those of the JCPDS reference card, so it is not needed to present the test data. The microprobe tests showed high Zn content, therefore we consider the sample a Zn-olivenite (Table 12).

Tyrolite

Tyroloite is the rarest arsenate mineral of the locality. It forms thin blue crusts in the fissures of calcite. The EDX examination proved the presence of Ca, Cu and As in the sample. The below X-ray reflections are convincing for tyrolite. The strongest peaks are: 27,7 (100), 13,7 (80), 2,98 (70).

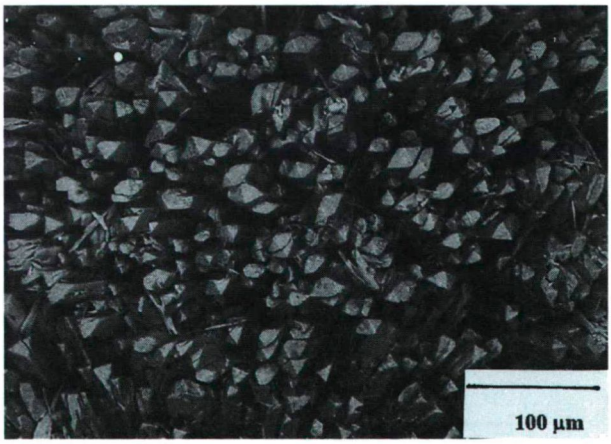

Fig. 10. Olivenite, columnar crystals, Pécs, Kozári vadászház quarry. Scanning electron micrograph.

TABLE 12

Chemical composition of olivenite from Pécs (Wt%)

CuO	37.71
ZnO	14.99
Fe_2O_3	–
Al_2O_3	–
P_2O_5	0.17
As_2O_5	39.69
Total	92.69

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

KŐVÁGÓSZŐLŐS

In the sedimentary, infiltrated uranium ore deposit rarely arsenate minerals have been observed in dolomite concentrations. At first VINCZE (1979) found tennantite that must have been the primary As containing mineral. We ourselves found 1–2 mm large tetrahedral crystals of tennantite in charred wood trunks.

Picropharmacolite and *talmessite*

These arsenates were found in the cavities of dolomite concentrations accompanied by calcite and dolomite. They form white crusts and radiating aggregates on the surface of calcite crystals. For they formed close together the presence of both minerals is supposed merely by X-ray diffraction tests. The test made on the sample taken from the radiating masses provides good correspond with JCPDS card. Most important peaks appeared on the X-ray diffraction record (Table 13).

TABLE 13

X-ray powder diffraction data of picropharmacolite from Kővágószőlős

Picropharmacolite Kővágószőlős		Picropharmacolite JCPDS 14-0222	
d(Å) obs.	I/I_0	d(Å)	I/I_0
13.47	100	13.50	100
9.25	36	9.20	60
		.	
		.	
		.	
3.19	30	3.18	80
3.05	32	3.06	60
2.92	21	2.92	50

Made in ALUTERV-FKI (Budapest).

The microprobe analysis of this structure (Table 14) revealed that its chemical composition corresponds well with picropharmacolites from Richeldorf and Freiberg (DANA 1920), but not with talmesite analyses (BARIAND and HERPIN 1960). Based upon these data we consider the radiating aggregates as picropharmacolite. The SEM images clarified that the crystals are actually of lamellar habit (*Fig. 11*).

TABLE 14

Chemical composition of picropharmacolite from Kővágószőlős, (Wt%)

MgO	2.92
CaO	29.05
BaO	0.08
Al_2O_3	1.87
As_2O_5	48.60
Total	82.52

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

Fig. 11. Picropharmacolite, needles crystals, Kővágószőlős. Scanning electron micrograph.

The X-ray diffraction tests made on the sample taken from the powder-like crusts identified talmessite (Table 15). 21 peaks appeared on the record in godd correspondance with its reference card. Quantitative analysis has not been made so far, its qualitative test indicated Ca, Mg, As elements which are characteristic to talmessite.

MÁD

North of the village in a limnoquartzite outcropt 1 cm thick alunite veins fill in the fissures of quartzite and several mm large alunite crystals occur in the cavities of the

TABLE 15

X-ray powder diffraction data of talmessite from Kővágószőlős

Talmessite Kővágószőlős		Talmessite JCPDS	
d(Å)	I/I_0	d(Å)	I/I_0
6.468	w	6.43	5
5.149	vs	5.09	60
4.665	s	4.62	40
4.286	vvw		
4.004	vv	3.95	10
3.789	vvw		
3.597	m	6.56	60
3.341	m	3.34	40
3.225	s	3.21	80
3.101	s	3.07	100
3.014	w		
2.796	vs	2.77	100
2.611	w	2.67	5
		2.58	20
		2.51	5
2.465	vw	2.44	20
2.355	vvw	2.32	5
2.177	w	2.16	40
2.132	vvw	2.12	5
		2.07	10
2.040	vw	2.04	5
		2.01	20
1.905	vw	1.89	20
		1.829	20
1.800	vw	1.782	20
1.733	w	1.717	60
1.709	w	1.697	40
1.661	vvw		
1.601	vvw	1.608	5
1.572	vw	1.584	10
		1.538	20

vs→vvw: very strong → very weak
Made in British Museum Natural History, London

veins. Scorodite and pharmacosiderite were rarely been observed on the alunite crystals. Unlike the other arsenate localities, we have not found any sulphide so far can be considered the arsenic bearing primary sulphide. Only pyrite was found, but contained no As, not even in enclosures based upon the microprobe analyses. It has long been known however that As is a characteristic trace element in the limnoquartztes of the Tokaj Mts., so the presence of arsenates is not surprising (VETŐ 1971, SZAKÁLL, 1990).

DOBOSI (1980) found tetrahedrite in the neighbouring area (Király Hill) so it is possible that is could have also been formed in the right next environs of the locality in question.

Scorodite

It is most often found as light green yellow or brown crusts and spherical aggregates in the fissures, reaching some mm, built up from short, columnar crystals. Bypiramidal crystals were also found (*Fig. 12*) with the (111), (201), (120) planes on them (*Fig. 13*). The crystals are 0,X mm large, light green coloured. The X-ray examinations assured the corresponding with those of the ICPDS card. The quantitative microprobe analysis also proves the presence of scorodite (Table 16).

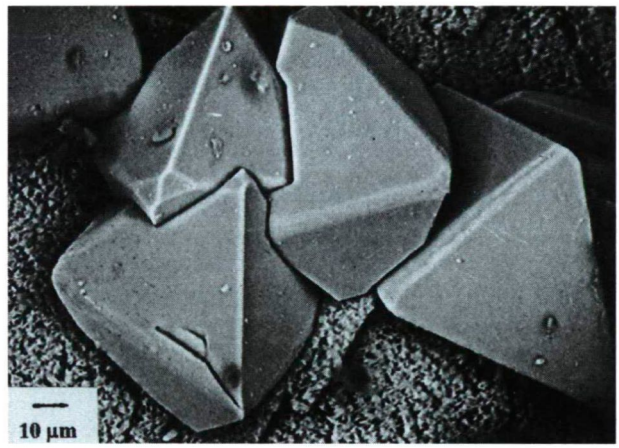

Fig. 12. Scorodite, bypiramidal crystals, Mád. Scanning electron micrograph.

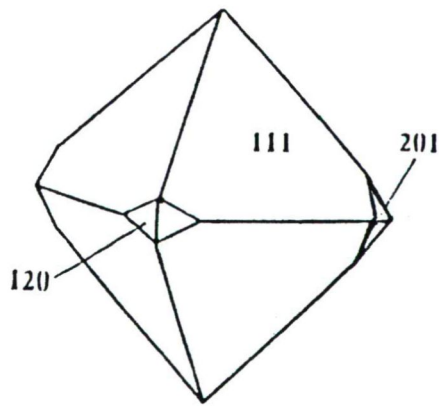

Fig. 13. Crystal drawing of scorodite from Mád

TABLE 16

Chemical composition of scorodite from Mád, (Wt%)

Fe_2O_3	33.12
MnO	0.10
Al_2O_3	0.51
CuO	0.13
As_2O_5	50.15
Total	84.01

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

Pharmacosiderite

It is found in the fissures of quartzite accompanied by scorodite. It forms light green or yellow hexahedral crystals (*Fig. 14*). There were 25 peaks observed on the X-ray record showing good correspondence with the reference card (Ba-pharmacosiderite) (Table 17). The only major peak that did not appear on the record is the one at 8.08 Å. According to the quantitative microprobe analyses there are two characteristic compositions, one of them is poor in K, the other is shown on Table 18. Ba-pharmacosiderite was first described by WALENTA (1966) as a newly discovered mineral, but it was later discredited by the IMA comitte. PEACOR and DUNN (1985) studied numereous Ba containing pharmacosiderites: comparing them with that in question the Ba/K ratio is nearly identical, the Fe content is similar. Therefore the mineral in queastion is Ba containing pharmacosiderite.

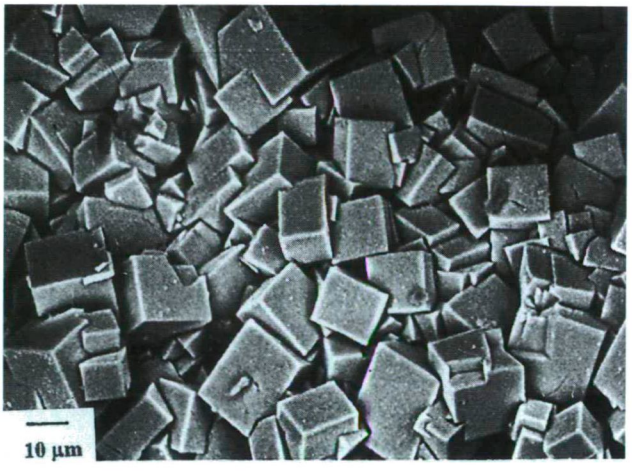

Fig. 14. Pharmacosiderite, cube-like crystals, Mád. Scanning electron micrograph.

TABLE 17

X-ray powder diffraction data of pharmacosiderite from Mád

Pharmacosiderite Mád		Pharmacosiderite – Ba JCPDS 34-154	
d(Å)	I/I0	d(Å)	I/I0
7.98	91	8.08	80
		7.95	95
		5.68	5
5.62	11	5.64	5
4.61	32	4.619	40
4.019	25	4.031	30
3.980	46	3.980	40
3.573	23	3.565	40
3.272	60	3.282	30
3.254	100	3.256	70
2.8241	80	2.833	100
		2.814	55
2.693	24	2.691	10
		2.6705	11
		2.6585	55
2.6540	38	2.6545	40
2.5727	20	2.5482	45
2.5336	40	2.5217	40
2.5164	48	2.5183	55
2.4458	58		
2.4182	31	2.4275	30
2.4009	46	2.4032	45
2.2910	?	2.3086	80
		2.2305	5
		2.1471	5
		2.1366	5
2.1302	9	2.1312	10
		1.9919	5
1.9520	10	1.9563	8
1.9460	5		
1.9388	14	1.9316	8
		1.8994	3
		1.8901	5
1.8749	28	1.8788	40
		1.8773	5
1.8306	12	1.8388	10
		1.8283	8
1.7939	22	1.8010	25
1.7815	30	1.7862	25

Made in Steiermärkisches Landesmuseum Joanneum, Graz

TABLE 18

Chemical composition of pharmacosiderite from Mád (Wt%)

K_2O	0.05
BaO	9.92
SrO	–
CaO	0.03
Fe_2O_3	39.37
Al_2O_3	1.91
P_2O_5	0.07
As_2O_5	43.26
Total	94.61

Analyst: BIRCH W. D. (Museum of Victoria, Melbourne)

GYÖNGYÖSSOLYMOS

The first description of the hydroquartzite with Sb traces on Asztagkő Hill was made by SZUROVY (1940). From the alteration of the Sb bearing primary minerals in a brecciated zone, a diverese secondary mineral association was formed. DOBOSI (1979, 1982) discovered members of the tetrahedrite group of various chemistry by microprobe tests made on samples from drillholes sunk in the neighbouring area.

Pharmacosiderite

In course of the examination of the Sb bearing mineral association, light yellow hexahedral crystals were observed. The surface of this very rarely found mineral reaching 2 milimeters in rugged, uneven as shown on the SEM images. By X-ray tests they turned out to be pharmacosiderite. According to its qualitative microprobe analysis the quantity of Ba and K are nearly the same.

ACKNOWLEDGEMENT

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GEOCHEMICAL CHARACTER OF AMPHIBOLITES FROM TISZA UNIT ON THE BASIS OF INCOMPATIBLE TRACE ELEMENTS

M. TÓTH TIVADAR*

Department of Mineralogy, Geochemistry and Petrology Attila József University

ABSTRACT

Pre-metamorphic origin of amphibolites and amphibolic gneisses from Kőrös Complex (Eastern Tisza Unit) has been an unsolvable problem for a long time. Last year the trace element (Zr, Y, Nb, Ti, P) composition of almost 100 amphibolite samples were measured and this data base may be useful for determining pre-metamorphic rock types and petrochemical character of amphibolites. Present examinations show that – opposite to the previous results – there were no alkaline magmatites in the area, but the rocks of both tholeiitic and calk-alkaline series were formed. The data also suggest that the original igneous rocks of amphibolic gneisses developed in an island arc, while the amphibolites' either in a fore-arc or in a back-arc basin.

INTRODUCTION

The best and – by numerous drillings – the most densely recovered district of the Tisza Unit's crystalline basement is a part of the Kőrös Complex in the "Tiszántúl" region (previous Kőrös and Szeghalom Formation (SZEDERKÉNYI 1984), or Kőrös–Berettyó Unit (BALÁZS ed. 1984)). This area is approximately bounded by the river Kőrös, the southern boundary of the Szolnok–Máramaros flysch zone, the river Tisza, and the eastern border of Hungary (*Fig. 1.*). Due to the relative many deep drillings this is the most promising district of the whole Tisza Unit from the point of view of modelling metamorphic and pre-metamorphic evolution.

Numerous attempts have been made to explore the paleotectonic position previous to the presumably Variscan metamorphism primarily by examining the mica-schists, gneissose rocks constituting the majority of the crystalline complex and the amphibolites forming thin intrcalations from the point of view of the geochemical nature of their major and partly of their trace elements (e.g. SZEDERKÉNYI 1980, 1983, BEDINI *et al.* 1993.)

Having examined the gneisses and mica-schists the data for the original quality of rock indicate a one-time greywacke-pelitic type sediment (SZEDERKÉNYI 1984). Consequently, a model of sedimentation in eugeosyncline and, in relation, ophiolitic, maybe island arc magmatism seems possible (SZEDERKÉNYI 1981). This theory is not supported undoubtedly by the analysis of the amphibolitic rocks. It is certain only that the majority must have been volcanic rocks, however, the possibilities of the exact type range from comatiite-like rocks to those of calc-alkaline, tholeiite and alkaline affinity. There have also been many attempts made to reconstruct the original tectonic setting.

* H–6701 Szeged, P.O. Box 651, Hungary

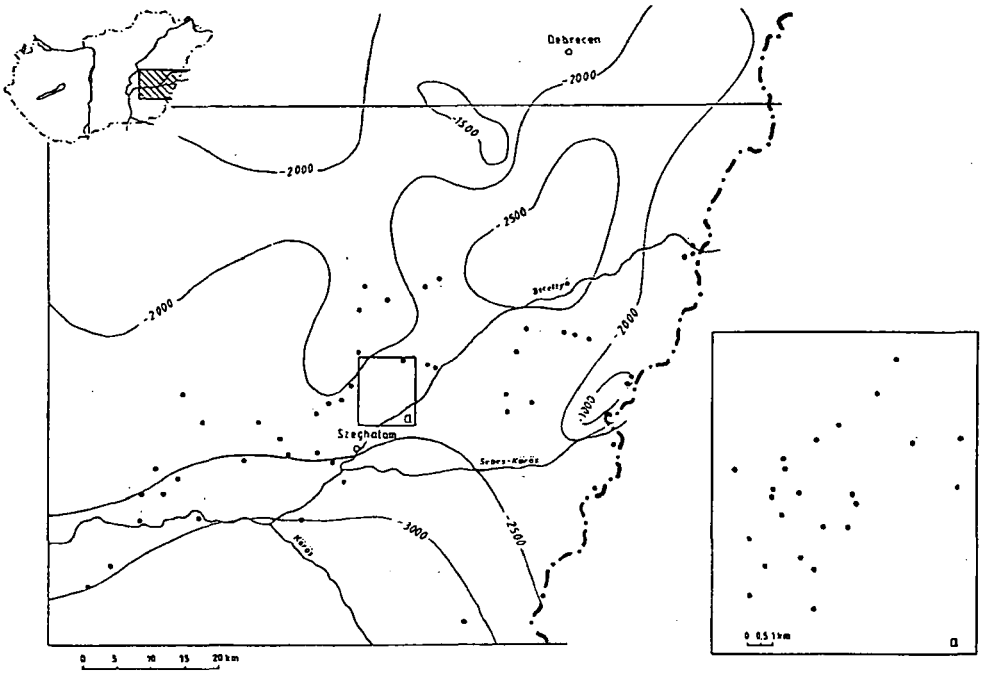

Fig. 1. Distribution of the examined area in Hungary and the locality of the boreholes.

It is evident that the given rock types and series can't have developed an the same time in any place. Therefore the creation of a common model from the results is rather complicated, if it is possible at all. The cause of the problem maybe the fact that during the previous studies only the geochemical data of the major elements were available to the authors and even those not in a required number. And an exact result cannot be expected from data of single samples in the case of metamorphic, what is more, poly-metamorphic rocks.

Recently trace elements have been measured from more than 100 amphibolite and amphibolic gneiss samples. The measurements were taken by X-ray fluorescence analysis at the Institute of Geosciences of the Johannes Gutenberg University in Mainz, Germany. Due to the large number of analyses of the incompatible trace elements (P, Ti, Y, Zr, Nb) considered also immobile, an opportunity presents itself to reconstruct, more precisely than was done before, the geochemical features of the former magmatic rocks and the tectonic position of their development.

GEOLOGICAL SETTING

The metamorphic body of the Kőrös Complex is formed mainly by medium and high grade metamorphic rocks, first of all by gneiss an mica-schist. In its axial zone lies a migmatite and a narrow granite belt. Into this major crystalline complex thin (maximum 3–4 metres thick) amphibolite and rarely leptinite bodies are intercalated.

The area has developed by the result of a multi-stage metamorphism. The most significant one of those was a Barrow-type amphibolite facies metamorphism, which at places exceeded even the sillimanite isograd. By a K/Ar chronology its age could be estimated to about 320–330 million years. The uplift after the Variscan orogeny was accompanied with considerable erosion and retrogressive processes. The last major metamorphic event was linked with the Alpine orogeny, when mainly different tectonites were developed and in the joint systems hydrothermal mineral association (laumontite, calcite, chlorite, pyrite) was formed due to lateralsecretion.

PETROGRAPHICAL FEATURES OF THE AMPHIBOLITES

Petrologically, the selected samples show a great variety, though the majority consist of the minerals of the general amphibolite paragenesis (e.g. LAIRD and ALBEE 1981). All samples contain acid-intermediate plagioclase, hornblende an quartz. Most samples contain epidote (zoizite, clinozoizite), chlorite, Ti-phase (mainly sphene), biotite and apatite. The chlorites are exclusively of retrograde origin, relics showing a progressive direction cannot be found. Hornblende is occasionally replaced by actinolite and/or bluish-green barroisite type amphibole. In some samples pumpellyite, prehnite and garnet appear indicating the main steps of metamorphic evolution. Rarely scapolite and zircon can be found as relic igneous minerals. The amphibolites do not contain any other relic mineral grains or textural marks suggesting magmatic origin. Although accurate mineral quantification from thin sections has not been taken, on the basis of the quantity of the rock-forming minerals two major groups can be marked out.

The rocks belonging to the first one are massive, greenish-black amphibolites with mainly nematoblastic texture almost entirely made up of amphibole and, in minor part, plagioclase. They often contain sphene, ilmenite, garnet, epidote as accessory minerals. The samples of the other group consisting of less amphibole and more feldspar can be considered as amphibolic gneiss. Here biotite and quartz are also recurrent components. The difference is unexplainable with metamorphic reasons, it may have been caused by the different original rock quality.

GEOCHEMICAL CHARACTER OF THE AMPHIBOLITES

Variables applied

In the course of different metamorphic processes the major elements and a vast majority of the trace elements may mobilize. Therefore however useful these elements could be for classification of fresh igneous rocks and for the identification of their chemical feature, they cannot be applied unboundedly for the geochemical analysis of amphibolites. Ti, Y, Zr, Nb and P however, may suit this requirement.

– Because of their high charge/radius ratio they are transported in aqueous fluids only in an exceptional case (e.g. with high F^- activity) (PEARCE and NORRY 1979). As a result they can be regarded immobile also in different metamorphic conditions (e.g. GRAHAM 1976, Pearce 1975, WOOD *et al.* 1976, COISH 1977, WINCHESTER and FLOYD 1984). The incidental mobility of Zr and Ti may be caused by the high CO_2 content of water moving inside the rock (MORRISON 1978), while contact with sea-water could

make Ti mobility again (MIYASHIRO 1975). In metamorphic conditions an increase in P-content can often be observed, too.

– On the other hand the change of their quantitative ratios presents each tectonical position conspicuously, and in most cases they characterize precisely the different magma types (PEARCE 1975, PUGLISI *et al.* 1988).

– Finally, in the course of basalt fractionation mainly these elements show incompatible behaviour (COX *et al.* 1979), though in case of a possible crystallization of amphiboles and Fe-Ti oxides the Ti-content of the residual melt decreases suddenly, and so the titanium cannot be regarded incompatible (MIYASHIRO an SHIDO 1975, ALABASTER *et al.* 1982). In the case of clinopyroxene crystallization the distribution coefficient of yttrium might be considered high (0.5–4) (MANN 1983).

Owing to these characteristic features, with proper care, the given element group can be suitable for modelling the pre-metamorphic evolution of metamorphosed basalts and andesites.

Study of immobility

The examination concerning the tectonic setting of rocks must be preceded by proving the immobility of the elements to applied. There are several ways to approach this question. The most obvious one is to set out from the correlation system of elements (EVANS *et al.* 1981). As incompatible elements are in question, it is probable that in the original rocks the correlation coefficient values were significant. These values might decrease considerably in case of mobilization. Therefore a significant correlation would indicate immobility, while the lack of it would indicate mobility. On the other hand, due to similar solutional characteristics, these correlations may last after mobilization processes (RIVALENTI and SIGHINOLFI 1969). Though the values of the correlation coefficient are not high (Table 1.) at the given level (0.99) all values [except r(Nb, Y)] are significant.

TABLE 1.

A correlation matrix of examined elements

	Zr	Y	Nb	Ti	P
Zr		0.59	0.61	0.49	0.64
Y			0.20	0.37	0.57
Nb				0.71	0.62
Ti					0.67

Based on these results the relative immobility of the five elements can be supposed, and they can be used further on for specifying the origin of the amphibolites on the basis of the methods known in igneous geochemistry.

Geochemical features of the samples examined

As the elements examined can be used successfully for solving geochemical problems a number of methods have been set up for their usage. Most of them are discrimination diagrams applying two or three variables. Besides ratio diagrams and

spidergrams can also be found. These methods may answer three main problems connected closely with each other: the rock-type, its affinity (e.g.: tholeiitic) and the proper tectonic setting of development (e.g.: **WPB, MORB**). Diagrams generally deal with these questions with some overlapping, very often the same method is expected to define the chemical features of a certain rock and the tectonic setting of its development. On the other hand the complementary categories do not always appear together (e.g.: tholeiitic-calc-alkaline), consequently only a negative result can be accepted with absolute certainty. Because of the same reason a well selected set of methods should be applied, and their correct sequence may result an effective system (SZEDERKÉNYI 1983).

The first task is to determine the type of the rock. The graduation elaborated by FLOYD and WINCHESTER (1977, 1978) uses the today valid IUGS terminology (LE MAITRE 1979) and is easily reconcilable with it. Similar to the **TAS** system a variable responsible for alkalinity (Nb/Y) and one for fractionality (Zr/TiO_2) are used. The importance of the latter is emphasized by ALABASTER *et al.* (1982) and using the Nb/Y ratio as alkalinity index is widespread in literature (for the first time PEARCE and CANN 1973).

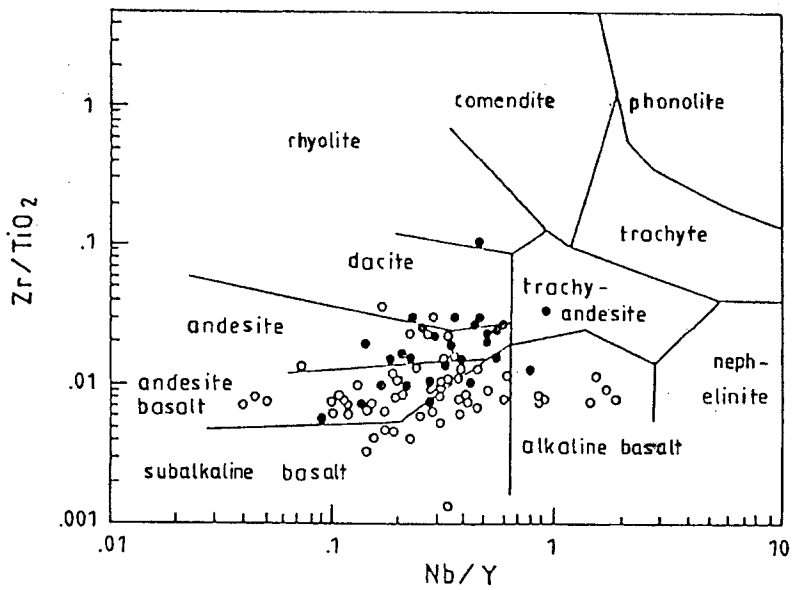

Fig. 2.
Nb/Y vs. Zr/TiO_2 (WINCHESTER and FLOYD, 1977) diagram for determining the rock types of amphibolites

According to the diagram most of the studied amphibolite samples are subalkaline basalts, basaltic andesites and some of them are andesite (*Fig. 2*). The amount of samples in the alkaline basalt field is not significant. On the basis of the two other diagrams (FLOYD and WINCHESTER 1975) used for separating tholeitiic and alkaline basalts it can be taken for certain that the presence of alkaline basalts can be left out of account. (*Fig. 3.*). Owing to the high potassium-content of the amphibolites in some

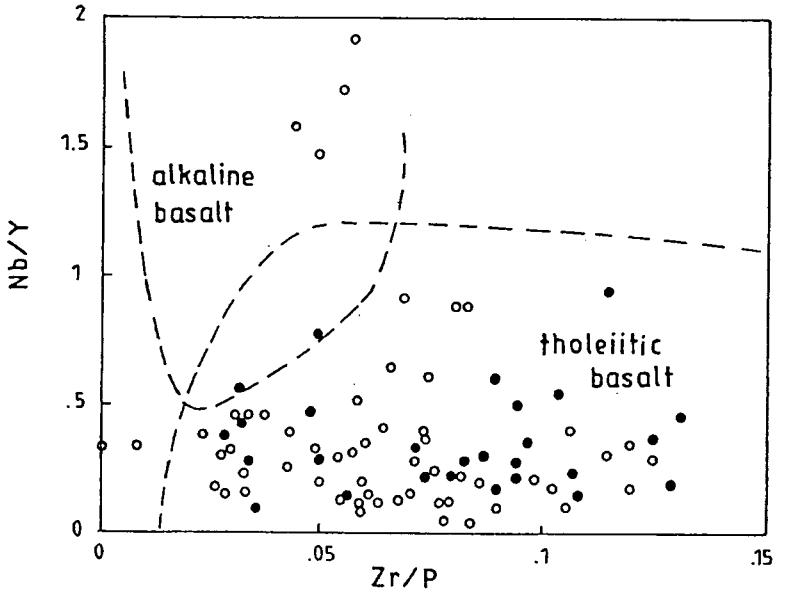

Fig. 3.
Zr/P vs. Nb/Y (FLOYD and WINCHESTER, 1975) diagram for characterizing alkaline and tholeiitic type basalts

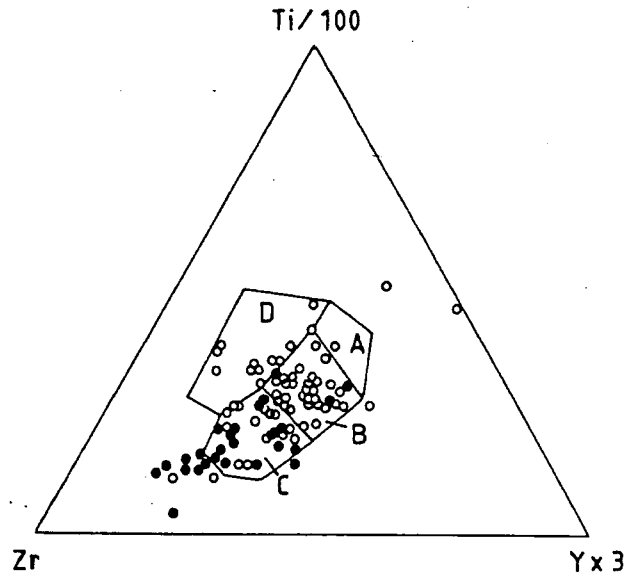

Fig. 4.
Ti-Zr-Y (PEARCE and CANN, 1973) diagram for classifying basalts – **A, B:** low-K tholeiite; **B, C:** calc-alkaline basalt; **B:** ocean-floor basalt; **D:** within-plate basalt

of the samples earlier the alkaline basalt type also seemed possible (SZEDERKÉNYI 1980, M. TÓTH 1993). The high alkaline-content, however, may also be a result of a later, alkaline metasomatic effect, maybe in connection with granitization, which did not influence the trace element composition considerably (SZEPESHÁZY 1971).

In the case of the diagrams above calc-alkaline basalt was not among the offered possibilities. That's why the tholeiitic affinity of the samples is not regarded to be proved for the moment. The problem is just the opposite with the Ti-Zr-Y classification (PEARCE and CANN 1973) where a calc-alkaline field appears at the expense of the tholeiitic one (*Fig. 4.*). Here – as a negative result – island arc (**IAB**) and withinplate (**WPB**) basalt can be excluded though the accuracy of this latter field has been in doubt (ZECK and MORTHOST 1982, HOLM 1982). These statements are verified by the results of the Ti-Zr (PEARCE and CANN 1973) and the Zr/Y-Ti/Y (PEARCE and GALE 1977) diagrams (*Fig. 5., 6.*).

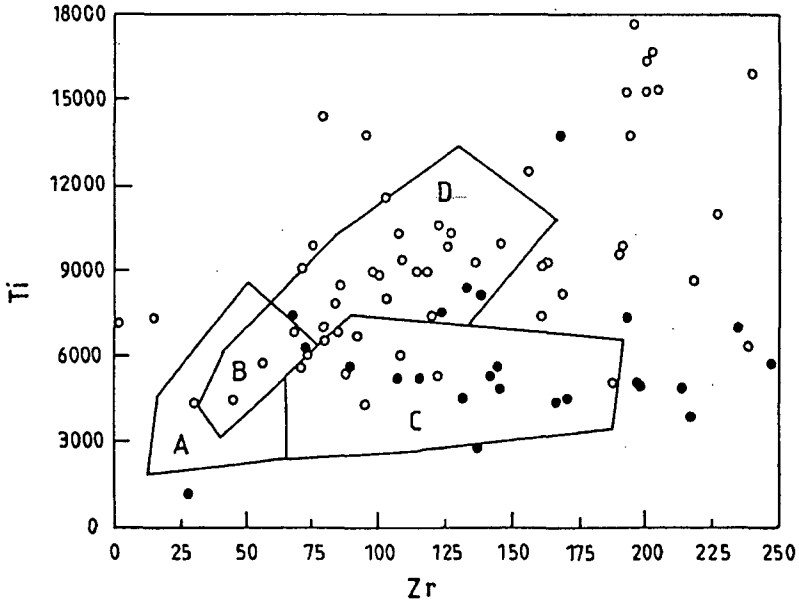

Fig. 5. Ti vs. Zr (PEARCE and CANN, 1973)
diagram – **A, B:** low-K tholeiite; **A, C:** calc-alkaline basalt; **B, D:** ocean-floor basalt

PALEOTECTONIC SETTING OF THE ROCKS

Relying upon all that have been said, it can be stated that the parent rocks of amphibolites must have been sea-floor, subalkaline basalts, andesites. To be able to determine whether the samples are from tholeiitic or calc-alkaline series and the tectonic setting of their origin further examinations are needed. According to the diagrams a genetic contact of the rocks with continental or within-plate systems can be excluded and formation of island arc is not possible either. Therefore among the petrotectonic positions the midoceanic ridge seems to be the most possible.

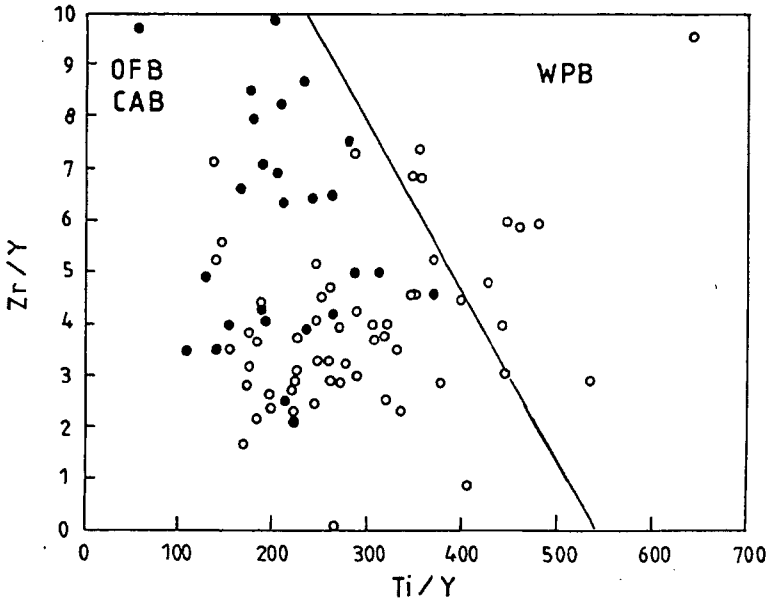

Fig. 6. Zr/Y vs Ti/Y (PEARCE and GALE, 1977) diagram

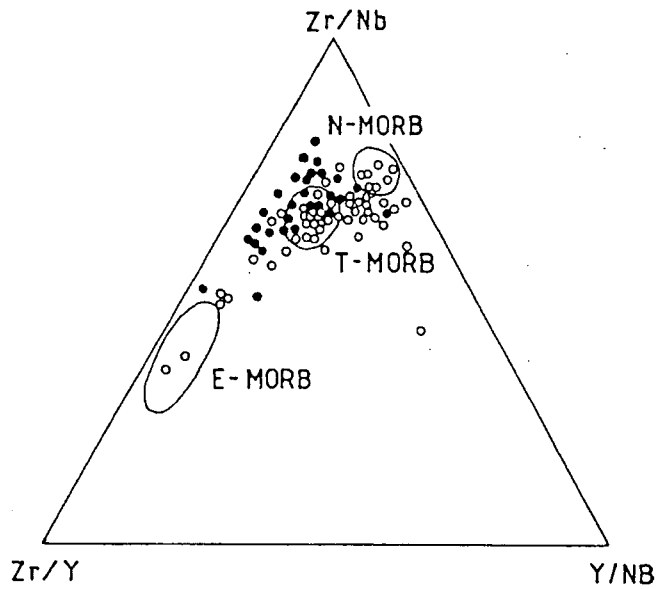

Fig. 7. Zr/Y-Y/Nb-Zr/Nb (FODOR and VETTER, 1984) diagram for distinguishing **MORB** basalts

The fact that in case of partial melting to various extent the five elements studied appear in the melt in different quantities (HESS 1989) makes the classification of **MORB**-type basalts possible and the conditions of development more exact. Especially the increase of Nb-quantity indicate well the enrichement (**E-MORB**) of the original substance (**N-MORB**) (LE ROEX *et al.* 1983). The development of transitional magma (**T-MORB**) can be explained by hybridization of the two previous types (LE ROEX *et al.* 1983). This change is shown clearly by the alteration of the Zr/Nb ratio, as Zr-content is not reactive to the changes above (WEAVER *et al.* 1979, SAUNDERS *et al.* 1988). On the basis of average value of the Zr/Nb ratio (18.6) in our case **T-MORB** type is possible (ERLANK and KABLE 1976).

The enrichment of the original melt can be followed if not a usual discriminating diagram but a ratio diagram is used. The elements applied are incompatible during both tholeiitic and calc-alkaline differentiation and their immobility can also be proved. Therefore it is probable that the value of their ratios will be close to constant, for by measuring the ratios the often disturbing effect of magmatic differentiation could be excluded (MYERS and BREITKOPF 1989). In this way on the Zr/Nb-Zr/Y-Y/Nb diagram the hybridization trend of rift basalts can be seen (*Fig. 7*) (FODOR and VETTER 1984). As the figure shows most sample are of **T-MORB** type and only some of them are of **N-MORB**. The number of rocks enriched with Nb (**E-MORB**) is subservient. The majority of the samples shows **T-MORB** character also on the Nb-normalized spidergram (MYERS and BREITKOPF 1989) (*Fig. 8*). All things considered it can be stated that the original magma went through a slight Nb-enrichment.

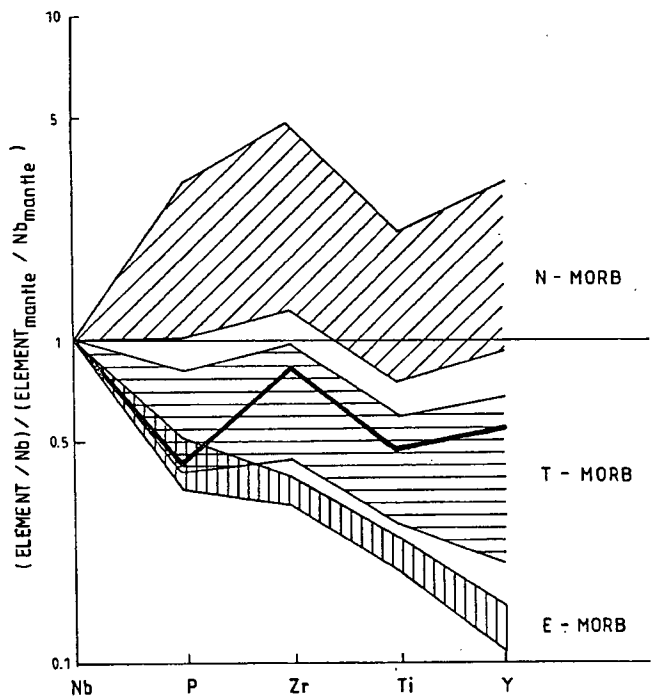

Fig. 8. Nb-normalized spidergram (MYERS and BREITKOPF, 1989; modified). Solid line represents the examined amphibolites

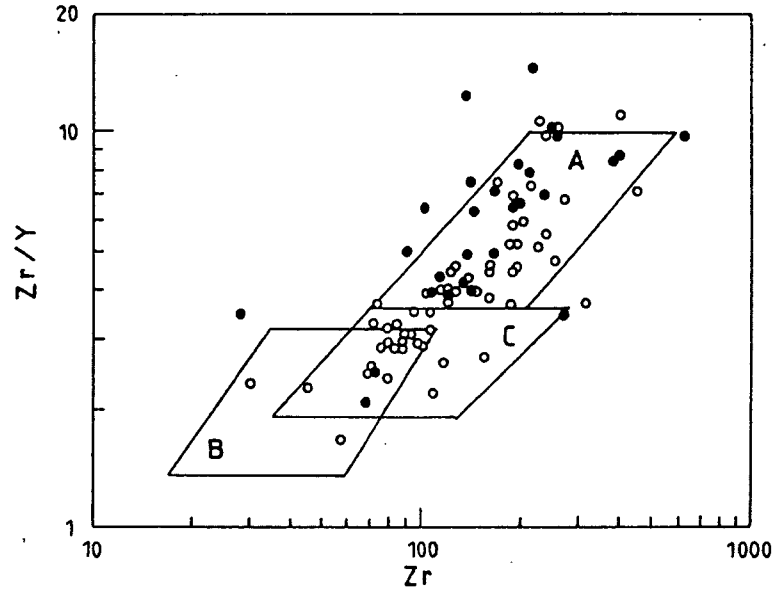

Fig. 9.
Zr vs. Zr/Y (PEARCE and NORRY, 1979) diagram for distinguishing basalts of the main tectonic setting – **A:** Within-plate basalt; **B:** Island-arc basalt; **C:** Mid-ocean ridge basalt

Mineralogically the main difference between the two subalkaline trends is that in the calc-alkaline one the crystallization of magnetite (generally Fe-Ti-oxides) and amphiboles plays an ipmortant part too (MIYASHIRO and SHIDO 1975, HESS 1989). As in these minerals the distribution coefficient is large (1.5; 7.5) (PEARCE and NORRY 1979), the Ti-content increasing together with Zr indicates a tholeiitic, while the decreasing one a calc-alkaline differentiation trend (BECCALUVA *et al.* 1977). Using this figure (*Fig. 5*) – in accordance with other diagrams – suggests that rocks of both series were present on the examined territory.

The simultaneous presence of both series – crossing what has been said – suggests an island arc development (MIYASHIRO 1975). This supposition seems to be supported by some other facts too which contradict development of rocks in a rift:

– the large (nearly 50 percent) ratio of rocks of andesitic composition connected almost exclusively with island arcs (BAILEY 1981);
– the greywacke type composition of original sediments;
– the absence of any trace suggesting ophiolite association;
– the little width and quantity of amphibolite bodies.

CONCLUSIONS

Why is it that all the geochemical methods applied up to this point unanimously reject the possibility of island-arc development?

With regard to the elements studied from the point of view of the island arcs the most characteristic features are negative Nb-anomaly (EMMET 1987, SAUNDERS *et al.*

1988) and low Zr-content (PEARCE 1975). The amphibolites in the vicinity of Szeghalom, however, show positive anomaly in the instances of both elements especially Zr, which can be regarded as a fact excluding the possibility of subduction. Therefore geochemically the origin connected with rift seems provable and that connected with island arc seems improvable, while relying upon the stratigraphic and geochemical character of the rocks nearby the former can be excluded categorically.

The elucidation of this contradiction is possible in several ways. The most apparent possibility is to assume the opening of a brack-arc basin parallel with the agency of the island arc. In this case the development of the new basin is characterized by the formation of rift-type tholeiites, while that of the island arc by the development of calc-alkaline rocks.

In sedimentary environment referring to the island arc rift type basalt may also appear in the territory of the fore-arc. Here the andesites having high content of Mg, Cr and Ni are of boninite type (JAKEŠ and MIYAKE 1984). Among the samples examined by us for three basaltic andesites Cr>800 ppm, Ni>400 ppm could be seen and the previous examinations have also shown samples with extremely high Mg content (SZEDERKÉNYI 1984).

As on every diagrams the main argument against formation on island arc was the significant positive Zr-anomaly, the explanation of it must be very important. It may arise that Zr became mobile independently from the other elements due to alteration of examined rocks. However this supposition is opposed to the results of the examination on immobility. Therefore already the original volcanics and also the magma may have shown positive anomaly. In case of continental subduction Zr often appears relatively high content far higher than that characteristic of the island arcs (JAKEŠ and WHITE 1972), and the Zr/Y ratio may be higher as well (BAILY 1981). This may be the reason why most samples in Zr-Zr/Y discrimination have proved to be of within plate origin. (*Fig. 9*)

Whether the oceanic side of the sontinental island arc is considered to be the place of volcanism or the existence of a back-arc basin is assumed, the majority of contradictions arising seems to be solvable. The final answer, however, can be given by the accurate examination of further trace elements and the rare earths of the amphibolites. The detailed geochemical analysis of the leptinites of acid volcanic origin reaching the quantity of the amphibolites in some places (SZEDERKÉNYI 1984) is also needed.

SUMMARY

Geochemical features and tectonic setting of development of the amphibolites and amphibole gneisses in the Kőrös Complex of the Tisza Unit has been a problematic question for a long time in the literature of the subject. Some, so far open, questions could be answered by the geochemical analyses of the basis of incompatible trace elements.

- Probably, pre-metamorphic volcanics of alkaline affinity cannot appear in the territory, the alkaline character of the rocks can be the result of an alteration effect.
- Relying upon subalkaline basalts and andesites the existence of both tholeiitic and calc-alkaline trends is proved, therefore volcanism on destructive plate margin is possible. On the basis of the thorough examinations development on continental margin is probable. The question is, however, whether the development of the tholeiitic basalts links with fore-arc or back-arc basin.
- The explanation to the significant Zr and less Nb anomaly is still not satisfactory.
- As, apart from some exceptions, the pre-metamorphic rock type of amphibolites has proved to be tholeiitic basalt, and that of the amphibole gneisses to be calc-alkaline andesite, it can be stated that the main reason for the petrological differences present among the examined metamorphites described before is the original quality of rock.

ACKNOWLEDGEMENT

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MAGMATIC FEATURES AND METAMORPHISM OF PLAGIOGRANITE ASSOCIATED WITH A JURASSIC MORB-LIKE BASIC-ULTRABASIC COMPLEX, BÜKK MOUNTAINS, HUNGARY

D. SADEK GHABRIAL,* P. ÁRKAI* and G. NAGY*

Laboratory for Geochemical Research of the Hungarian Academy of Sciences

ABSTRACT

Plagiogranite forms cross-cutting dikes, veins with sharp boundaries within the MORB-like Jurassic basic-ultrabasic complex (Szarvaskő complex) of the southwestern Bükk Mountains, NE-Hungary. Earlier petrological and geochemical interpretation suggests that the basic magmas became contaminated by terrigenous sediments in a shallow magma chamber (DOWNES *et al.* 1990), thus producing atypical, incomplete ophiolitic sequence. In order to characterise the formation of the plagiogranite in question, literature data on oceanic granitic rocks are critically reviewed. The Szarvaskő plagiogranite consists mainly of quartz, plagioclase, biotite ± garnet. Its major element bulk chemistry is similar to that of the oceanic plagiogranites (high Si, moderate Al, low total Fe+Mg and extremely low K contents), crystallised from late differentiated basic melt. The garnet being present in strongly varying proportions in granite is usually accumulated near or at the contact with the metaclastic enclaves. The garnet is almandine-rich, and displays slight chemical zoning of changing character. Most frequently, the Fe content increases and the Ca content decreases from the centres towards the rims of the grains. Its chemistry corresponds to that of the garnet found in metapelites or is plotted near to the field of the igneous silicic rocks in the diagrams used for discrimination. As the investigated metaclastic enclaves are represented by biotite-plagioclase hornfels, and garnet never forms contact metamorphic rim around the enclaves, but it is always dispersed in the granitic melt, the formation of the garnet is most probably a consequence of local assimilation of the metaclastic rocks at a deeper level. The magmatic minerals are extensively altered: the plagioclase is completely albitised, while the biotite is totally altered to chlorite, epidote, prehnite and titanite. The secondary mineral assemblage consists of quartz, albite, chlorite, epidote, prehnite, calcite, titanite, white mica (sericite) and smectite, while pumpellyite is lacking. Illite and chlorite "crystallinity" indices indicate anchizonal metamorphic circumstances, agreeing fairly well with the temperature data (270–350 °C, in average, 298 °C) obtained by the chlorite-Al^{IV} geothermometer of CATHELINEAU (1988). These data refer to metamorphic conditions corresponding either to the pumpellyite-prehnite-quartz or to the prehnite-actinolite facies. Earlier petrological data of ÁRKAI (1983) prove Cretaceous regional (dynamothermal) metamorphism in the surrounding formations. In order to get evidence on eventual preceding (ocean-floor hydrothermal or burial) metamorphic effects, detailed study of the whole ophiolite-like sequence is required.

INTRODUCTION

Complete ophiolitic sequences consist mainly of ultrabasic and basic rocks. Within these ophiolites, leucocratic rocks ranging in compositions from tonalite to albite granite are usually regarded as the final products in the crystallization of basic magma and they have generally been called keratophyre, quartz-keratophyre, tonalite, trondhjemite, quartz-diorite, albite granite, albitite or plagiogranite (WILSON, 1959, THAYER

* H–1112 Budapest, Budaőrsi út 45, Hungary

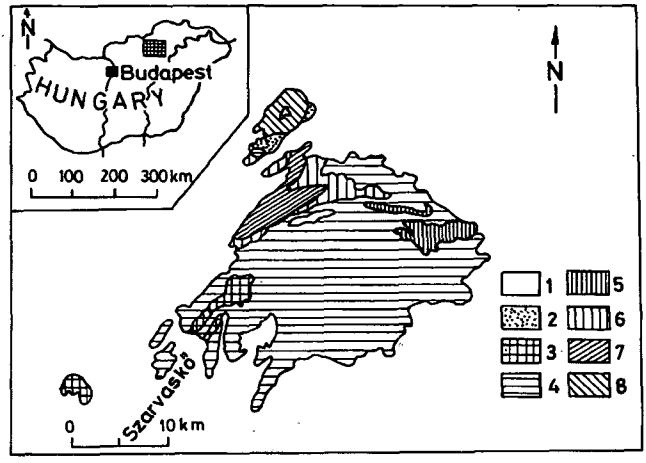


Fig. 1. Simplified geological sketch map of the Bükk Mountains (after BALLA 1982)
Legend: 1=Cenozoic, 2=Upper Cretaceous, 3=Jurassic basic and ultrabasic rocks (Szarvaskő complex), 4=Triassic and Jurassic sedimentary, 5=Triassic metavolcanic, 6=Permian sedimentary, 7=Middle-Upper Carboniferous sedimentary, 8=Devonian-Lower Carboniferous sedimentary formations metamorphosed in low temperature conditions

1963, COLEMAN and PETERMAN 1975, COLEMAN 1977). These acidic rocks frequently form a small portion in the ophiolitic sequences.

The innermost tectonic units of the Western Carpathians (South Gemeric Unit, Bükk Unit) contain dismembered fragments of Mesozoic ophiolitic suites. In one of their largest occurrences, i. e. in the Szarvaskő basic-ultrabasic complex of **MORB**-type (*Fig. 1*), a great variety of rock types (gabbrodiorite, diorite, plagioclasite, quartz-diorite, granite and their aplitic, pegmatitic and porphyritic variants) were described by SZENTPÉTERY (1953). The rocks with intermediate composition occur mainly as local differentiates ("schliers") within tha basic rock mass with continuous transitions towards the gabbro, while the acidic types form cross-cutting dikes, veins with sharp boundaries. SZENTPÉTERY (1953) also gave the modal and main element compositions of these rocks, the formation of which was explained by crystallisation differentiation.

In spite of the detailed petrographic descriptions of the various types mentioned above, the petrogenetic interpretation of these rocks are lacking. Publishing new rock textural, mineral paragenetic, mineral chemical and bulk chemical data, this paper presents contributions to the petrogenesis of the plagiogranite, with special references to the distinction of the magmatic and metamorphic processes.

NATURE OF GRANITIC ROCKS IN OCEANIC ENVIRONMENT

Mineralogically, the oceanic plagiogranites consist mainly of quartz and plagioclase. Ferromagnesian minerals occur in small amounts and are commonly represented by hornblende, pyroxene or biotite. K-feldspar is completely absent but present only in rare cases. Generally, the plagiogranite rocks can be distinguished from the similar continental leucocratic ones which derived from differentiation of mafic rocks by their

chemical composition. The former ones are characterized by high silica, moderate alumina, low iron-magnesium and extremely low potassium content (COLEMAN and PETERMAN 1975, COLEMAN 1977).

It is obvious that the oceanic leucocratic rocks usually show close relation with associated mafic rocks. Two types of boundaries between the granophyric rocks of Troodos Igneous Complex and the underlying gabbro were distinguished by WILSON (1959): the gabbro may be interdigitated with the more acidic rocks or it may be capped by massive granophyres which are rarely penetrated by gabbro apophyses, whereas the boundary with overlying diabase is generally gradual. Sharp boundary could be recognized by THAYER (1963) between the gabbro and the quartz dioritic rocks; the latters forming dikes, small pipes and irregular masses. Moreover, dikelets of Na-rich trondhjemite and quartz monzonitic composition cutting some granophyric diabase intrusions are described by ENGEL and FISHER (1975). BATLEY *et al.* (1970) referred to the occurrence of light-coloured pillow keratophyric lavas that are underlain by a volcanic succession including basalt and diabase. Furthermore, BATLEY and BLAKE (1974) noticed silica-rich extrusive and intrusive rocks: the finer keratophyre rocks occur as pillow lavas and as massive flows above the gabbro, while the coarser varieties occur as dikes or sills. In very rare cases, these silicic rocks show relationship with the ultrabasic rocks of ophiolites, however, as described by IWAO (1953) in COLEMAN and PETERMAN (1975) intrusions of silicic rocks are commonly present within serpentinized ultrabasic rocks.

Many attempts have been made to define the origin of oceanic leucocratic rocks. The early opinion had been hold by GILLULY (1933) who attributed the origin of albite granite to late-magmatic and post-magmatic metasomatism of an earlier quartz diorite. In fact, these leucocratic oceanic rocks have been shown by most investigators as a product of differentiation processes rather than metasomatic origin.

In the ophiolites of Troodos, the $^{87}Sr/^{86}Sr$ ratios of mafic varieties are similar to those of silicic ones. This similarity is consistent with the idea of low pressure differentiation of a parental subalkaline basaltic magma to produce the K-poor leucocratic rocks (COLEMAN and PETERMAN 1975).

Fractional crystallization experiments were performed on a primitive oceanic tholeiite at dry and 1 atmosphere total pressure by DIXON and RUTHERFORD (1979). In these experiments, the produced residual liquid is of basaltic composition, and with increasing crystallization (after 95% crystallization) the residual liquid became Fe-enriched basalt. This Fe-enriched basalt would be immiscible at a temperature of about 1010°C. A more Fe-enriched basaltic liquid and a granitic liquid which resemble in composition to the oceanic plagiogranites present in certain MOR regions were the immiscible phases produced. Therefore. silicate liquid immiscibility could be responsible for producing plagiogranites in some ophiolites. Furthermore, hydrothermal experiments were made by SPULBER and RUTHERFORD (1983) on a primitive oceanic thoeliite. After 90% crystallization the residual liquid is characterized by high silica and low K_2O contents, similar to low K_2O plagiogranites present in ophiolitic sequence. In these hydrothermal experiments, however, silicate liquid immiscibility did not take place.

GEOLOGICAL SETTING

In the southwestern part of the Bükk Mountains, NE Hungary, Mesozoic ophiolitic sequences include plagiogranites which occur in Tóbérc quarry near the village Szar-

vaskő (*Fig. 1*). These sequences are situated in the innermost part of Western Carpathians, and show close relation to the Dinarids and Southern Alps, too. At present these ophiolitic sequences and associated sediments are regarded as ancient parts of the Inner Dinaric belt that moved to their present position as a result of horizontal displacements of the Bükkium towards the east during the Tertiary (KOVÁCS 1982, BALLA 1984).

According to BALLA (1983) and BALLA *et al.* (1983), these ophiolites form nappe system. The internal structure of these nappes is very complicated with overthrusting and strong folding. Clearly the reconstructed original stratigraphic sequence of Szarvaskő begins with sandy terrigenous flysch sediments which are gradually replaced upwards by clayey sequences. Within this terrigenous sequence, thick intrusive sills ranging in composition from ultrabasic to acidic (including plagiogranite rocks which occur as nests in the intrusive gabbro) are located in the lower part. As regarded by BALLA (1984), these intrusive bodies can be compared with the sheeted dykes of a complete ophiolitic succession. The effusive rocks are situated in the upper part of the sequence and are represented by pillow lavas formed under deep-water conditions. Based on main and trace element (including REE) geochemical and Sr and Nb isotope data of DOWNES *et al.* (1990) suggested that the basic magmas of the complex were probably originally **MORB**-like, but during evolution in a shallow magma chamber, they became contaminated by terrigenous sediments.

As regarded by ÁRKAI (1983), the basic rocks of Szarvaskő were affected by pumpellyite-prehnite-quartz facies regional metamorphism, while the illite "crystallinity" and vitrinite reflectance data of the surrounding sedimentary rocks refer to low-temperature (anchizonal) metamorphic and late-diagenetic transformations. According to ÁRAKAI (1983) the age of the regional metamorphism is Alpine (Cretaceous, pre-Senonian).

The presence of acidic rocks within ophiolites of Szarvaskő has been defined by some investigators. As shown by SZENTPÉTERY (1953), these rocks exist as nests within gabbro. The plagiogranite consists of plagioclase, biotite and quartz with garnet (EMBEY-ISZTIN *et al.* 1985). Also DOWNES *et al.* (1990) reported that these rocks consist mainly of albite + quartz + biotite, with apatite, zircon, sphene and xenotime being present as accessories. Secondary minerals include chlorite, prehnite, actinolite and sericite. Garnets were observed in some quartz diorite, having low Ca- and high Fe-content indicating low pressure metasomatic origin (EMBEY-ISZTIN *et al.* 1985). The chemistry of these rocks show strong LREE. LILE and Th enrichment and they are strongly depleted in Ti, P and Sr relative to the incompatible elements Zr, Nd and Ce (DOWNES *et al.* 1990).

Contradiction can be found in discussion regarding the origin of plagiogranites within ophiolites of Szarvaskő. As stated by SZENTPÉTERY (1932, 1938a, 1938b, 1939a, 1939b, 1940, 1953) the Szarvaskő rocks were generated by differentiation which went in two directions, one giving way to ultramafics and the other to intermediate acidic rocks. This opinion has been modified by BALLA *et al.* (1983) who related their origin not only to differentiation process but also at least to progressive assimilation of the country rocks. Later BALLA (1984), using the analytical data of DOBRETSOV, reported that the garnetiferous varieties were formed by melting of sedimentary or metamorphic rocks while the garnet-free ones by assimilation of this molten material in the gabbro magma. EMBEY-ISZTIN *et al.* (1985) attributed to the garnet present in these plagiogranites neither xenocrystal nor cognate origin. As chemistry and zoning pattern of garnet strongly vary from grain, garnet might form by metasomatism under different conditions (EMBEY-ISZTIN *et al.* 1985).

METHODS

The plagiogranite samples were collected from the Tóbérc quarry and its surroundings located south to the village Szarvaskő (southwestern part of the Bükk Mountains).

The rock samples were examined macroscopically in order to define the mineral composition, and the micro-textural features, with special regards to the distinction of the primary and secondary minerals.

The XRD analysis was used to determine te modal composition, illite "crystallinity" (abbreviated as IC(002) which expresses the half-height width of the illite first 10 Å basal reflection; see KÜBLER 1968, 1990) and chlorite crystallinity, ChC(001) and ChC(002) which are the half-height width values of the chlorite 14 Å and 7 Å reflections, respectively; see ÁRKAI 1991). To determine such parameters, the rock samples were crushed first in a jaw crusher, and then in a mortar mill (Type Pulversette 2, Fritsch) for about 3 min as mentioned by ÁRKAI (1982) to avoid the effects on the quantities of small (~20mm) grain size fractions produced or on the values of IC. These parameters were measured on desoriented powder of the whole rock, highly oriented fraction less than 2μm grain size which were separated from aqueous (clay-water) suspension onto a glass slide, using the method of KÜBLER (1975), and was dried at room temperature. Solvation of <2μm fraction samples was done holding the preparations in an atmosphere saturated with ethylene glycol at 80°C for 4 hours.

The samples were analyzed by a Philips PW–1730 X-ray diffractometer at 45 KV accelerating voltage and 35 mA amperage. The XRD instrument was used CuK_α radiation, proportional counter, divergence and receiving slits of 1°, graphite monochromator, goniometer speeds of 2°/min and 1/2°/min, and time constant of 2 sec.

In the laboratory of Geochemical Research, the IC measurements were calibrated with those of KÜBLER's laboratory where the anchizone boundaries are 0.25 and 0.42, by using standard rock slab series kindly provided by KÜBLER (see ÁRKAI 1983, 1991).

Microprobe analysis of rock-forming minerals was performed on carbon coated thin sections, using a JEOL Superprobe–733 electron microprobe instrument. The analyses were done at 15 accelerating voltage and 30 nA beam current using a beam size of approximately 5–10 μm. The wave-dispersive spectrometric quantitative analyses were done using Bence-Albee correction methods (BENCE and ALBEE 1968) and spessartite (Si, Al, Mn), olivine (Mg), hematite (Fe), wollastonite (Ca), albite (Na), spessartite (Mn), rutile (Ti) standards for the other minerals.

Whole rock chemical analysis of major elements has been made in the Geochemical Research Laboratory by Zoltán Wieszt, employing AAS Perkin Elmer 5000, using lithium metaborate digestion. Other methods, such as gravimetric for SiO_2 and loss-on ignition, permanganometric for FeO, spectrophotometric for SiO_2, TiO_2 and P_2O_5, and volumetric for CO_2 determinations, have been also used.

PETROGRAPHY

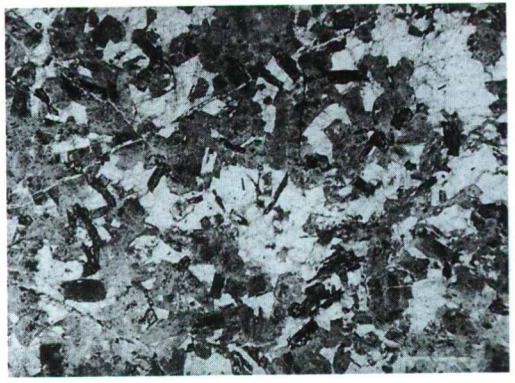
The plagiogranites of Bükk Mountains are fine to medium grained, light coloured rocks displaying typical granitic microstructure and composed mainly of quartz and plagioclase (*Plate I/1*), while K-feldspar has not been recognized in these rocks. The quartz occurs as xenomorphic crystals characterized average grain size ranging from 0.15 to 0.88 mm. Undulating extinction is found in some quartz crystals and notable cracks are also recognized, however, these cracks are sometimes filled with secondary

minerals such as sericite. Feldspar is commonly represented by unzoned, albitized plagioclase. According to the electron microprobe analyses, its composition varies between An_7 and An_1, with an average of An_2. The size of plagioclase may reach 2.34 mm in length and 1.08 mm in width, generally the average size ranges from 0.41 to 0.90 mm in length and from 0.17 to 0.38 mm in breadth. Certainly plagioclase crystals are highly corroded and cracked, and largely altered along cracks and cleavage planes or completely replaced by prehnite laths and sericite (*Plates* I/2 and *I/3*). Apatite and chlorite are found in plagioclase as inclusions. Biotite is the only primary ferromagnesian mineral present in minor amounts (*Plate I/4*). Biotite was replaced totally by secondary minerals (*Plates II/1, II/2* and *II/3*) including chlorite, epidote, prehnite (Fe-rich and Fe-poor) and titanite which in some crystals lies parallel to the (001) planes of chlorite. Inclusions of ilmenite, zircon, rutile, apatite and sphene are observed within altered biotite (*Plate II/4*). Other minerals, mostly ilmenite, pyrite, rutile, zircon and apatite are present as accessories. Garnet is frequently present in different proportions or my be completely absent. Depending on the abundance of garnet, the plagiogranites can be divided into two types, namely, garnet-poor or free and garnet-rich rocks (*Fig. 2*). The former type is usually found in Tóbérc quarry, containing gabbro rocks. It contains rounded, cracked garnet crystals (*Plate III/1*) ranging in size from 0.70 to 1.40 mm. These crystals contain inclusions of quartz, ilmenite, titanite and rutile. In the latter type coarse to fine grained garnets (average size reaches 0.35 mm) are commonly concentrated near to contacts with metasediment enclaves (*Plate III/2*). However, these garnet grains are embedded in the granite (*Plate III/3*), as plagioclase and quartz around garnet crystals commonly resemble those of the granitic rocks. These garnets contain mostly inclusions of quartz and ilmenite and their cracks are filled with titanite and prehnite. In general, colour zoning is absent in both garnet types as they exhibit little changes in their compositions from core to rim. The metasediments are usually associated with garnet-rich variety, and do not show preferred orientation. They seem to be xenoliths within acidic rocks, and consist essentially of fine grained, untwinned plagioclase and of chlorite that appears as a pseudomorph after biotite (*Plate III/4*). Monacite is present as accessory mineral. No quartz could be distinguished in these xenoliths. The plagioclase in these metasediments show prograde zoning: the Ca-content increases towards the edges. The contact between granitic rock and metasediment is sharp.

The fissure filling of these rocks (*Plate IV/1*) are represented by aggregates of chlorite, calcite, prehnite, quartz, albite, apatite and titanite. The iron content of prehnite in the aggregates is varying, the Fe-richer part is more or less located near fractures, and mostly forming exsolution laths (*Plate IV/2*).

Based on the XRD studies, the minerals present in the garnet-free or -poor plagiogranite of Tóbérc quarry are plagioclase, quartz and chlorite with minor prehnite and calcite. Traces of ilmenite, zircon, rutile, apatite, titanite, pyrite, epidote, kaolinite, smectite and garnet are also found. On th other hand, the minerals of the garnet-rich granites are mainly plagioclase, quartz, chlorite and garnet with minor calcite. Apatite, zircon, ilmenite, titanite, hematite, sphene, epidote, and prehnite are present as traces (Table I.) The measured XRD parameters are shown in Tables II and III.

The contact between granite and gabbro is more or less sharp. The average grains sizes of plagioclase and quartz decrease towards this contact (chilled margin).



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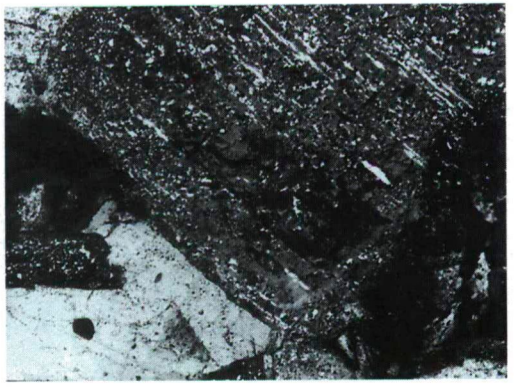

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4

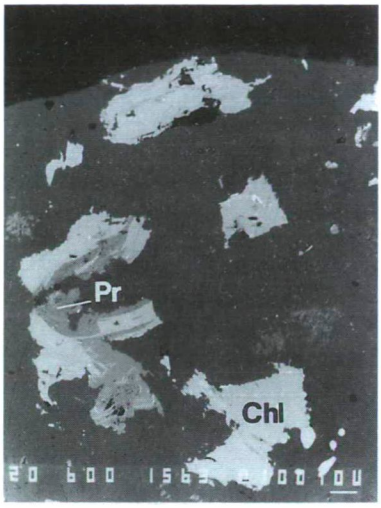
Plate I

1. Microtexture of plagiogranite (sample No. 1). Mineral composition includes quartz, plagioclase and chlorite. Fine cracks are filled with prehnite. Plane polarized light, approximate width of the photo is 6 mm.

2. Coarse plagioclase crystal altered along cleavage planes to sericite, while its cracks are filled with chlorite (sample No. 1). Crossed nicols, approximate width of the photo is 0.75 mm.

3. Laths of prehnite formed after plagioclase. Sample No. 5a. Crossed nicols, approximate width of the photo is 0.75 mm.

4. Chloritized biotite. Sample No 5c. Plane polarized light, approximate width of the photo is 0.75 mm.



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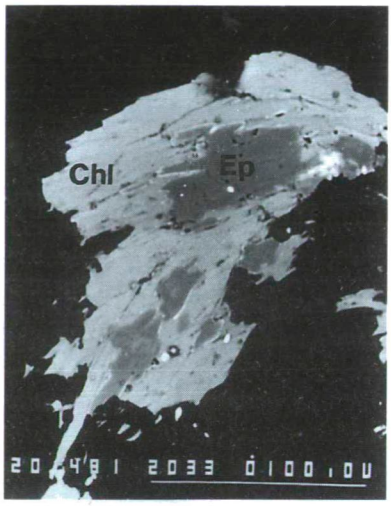

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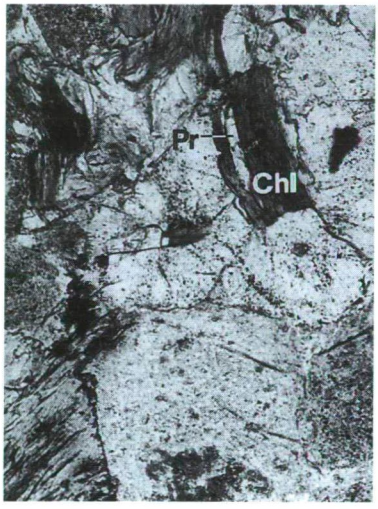

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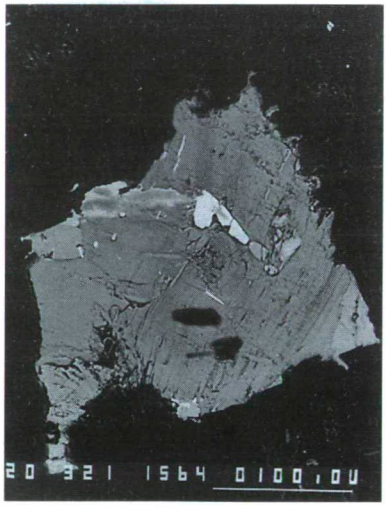
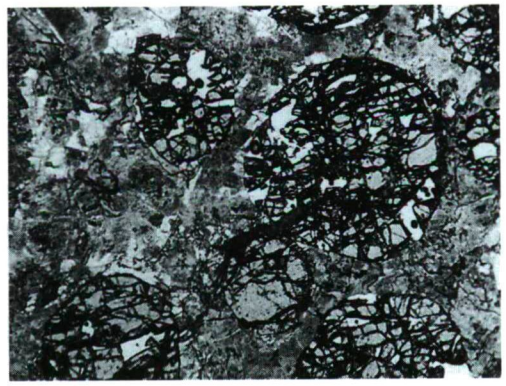


4

Plate II

1. Biotite altered to chlorite (Chl) and prehnite (Pr) and plagiclase to prehnite (Pr). Sample No 1. Back scattered electron (BSE) image, the length of the white bar in the lower right part of the photo is 0.1 mm.
2. Biotite crystal totally replaced by chlorite (chl) and epidote (Ep). Sample No. 34b. BSE image, the length of the white bar in the lower right part of the photo is 0.1 mm.
3. Biotite crystal altered to chlorite (Chl) and prehnite (Pr.). Sample No. 1. Plane polarized light, approximate width of the photo is 0.75 mm.
4. Inclusions of zircon and apatite within chloritized biotite. Sample No. 1. BSE image, the lenght of the white bar in the lower right part of the photo is 0.1 mm.

1



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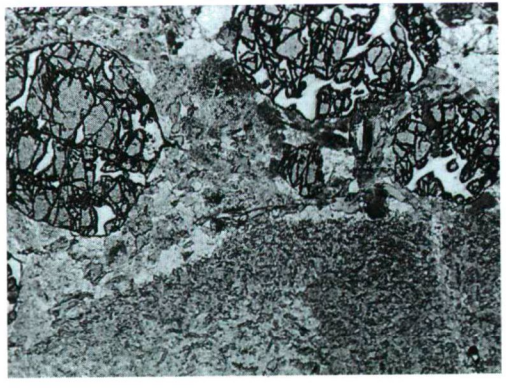

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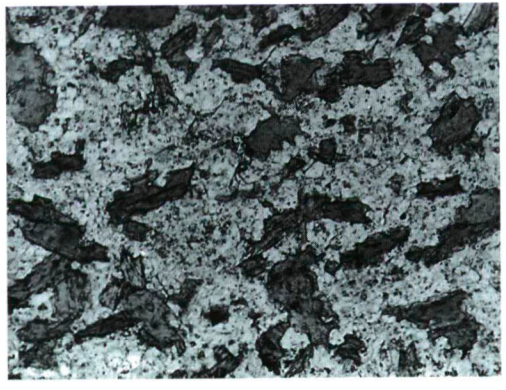

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Plate III

1. Garnet from garnet-poor granite. Sample No. 5b. Plane-polarized light, approximate width of the photo is 3 mm.

2. Abundant garnets from garnet-rich plagiogranite. Sample No. 34c. Plane-polarized light, approximate width of the photo is 6 mm.

3. Sharp contact between garnet-rich plagiogranite and metasediment enclave. Sample No. 34b. Plane-polarized light, approximate width of the photo is 6 mm.

4. Metasediment xenolith consists mainly of plagioclase and chloritized biotite. Preferred orientation is absent. Sample No. 34b. Plane-polarized light, approximate width of the photo is 0.75 mm.

Plate IV

1. Fracture filling with prehnite and calcite in garnet-poor plagiogranite. Sample No. 5c. Crossed nicols, approximate width of the photo is 3 mm,
2. Exsolution of Fe-rich prehnite in host Fe-poor prehnite. Sample No. 5b. BSE image, the legth of the white bar in the lower right part of the photo is 0.1 mm.

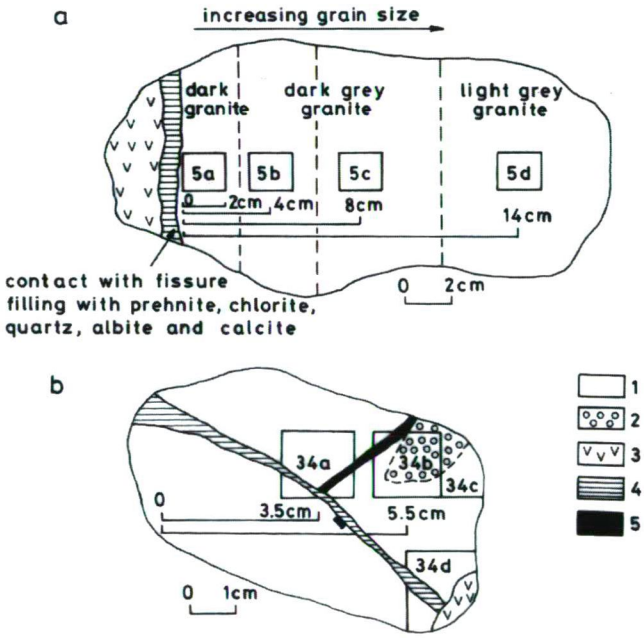

Fig. 2(a–b).

Sketch drawing of plagiogranite types of the Tóbérc quarry and the surroundings. 2a is a part of a garnet-poor granite dike showing the contact with gabbro. 2b is a garnet-rich granite also contacting gabbro but containing metasedimentary enclave, too.

Legend: 1=garnet-poor granite, 2=garnet-rich granite, 3=gabbro, 4=white fissure filling (prehnite, calcite, quartz), 5=metasedimentary xenolith

TABLE 1

Modal composition of the granites studied

Sample	Pl	Qtz	Chl	Ms	Alm	Smc	Kln	Prh	Pmp	Cal	Ep	Hem	Aln	Py	Ttn	Ilm	Zrn	Rt	Ap
1	+	+	x	tr	tr	tr	tr	o	–	–	tr	tr	tr	tr	tr	–	–	–	tr
5a	+	+	x	tr	tr	tr	–	o	–	o	tr	–	–	–	–	tr	–	–	tr
5b	+	+	x	tr	tr	tr	–	o	–	o	tr	–	–	–	–	tr	–	–	tr
5c	+	+	x	tr	tr	tr	tr	o	–	o	tr	tr	–	–	tr	tr	–	tr	tr
5d	+	+	x	tr	–	tr	tr	o	–	o	tr	tr	tr	–	tr	–	tr	–	tr
34a	+	+	x	tr	tr	tr	–	tr	–	tr	tr	–	–	–	–	–	–	–	–
34b	+	+	x	tr	+	tr	–	tr	–	tr	tr	tr	–	–	tr	–	tr	–	tr
34c	+	+	x	tr	+	tr	–	tr	–	o	tr	–	–	–	–	tr	–	–	tr
34d	+	+	x	tr	–	tr	–	o	–	o	tr	tr	–	–	tr	tr	–	–	tr
34e	+	+	x	tr	–	tr	–	tr	–	tr	tr	–	–	–	–	–	–	–	tr
35	+	+	x	tr	+	tr	tr	tr	–	–	tr	o	–	–	–	–	–	–	tr
36	+	+	x	x	–	–	–	–	–	–	tr	–	–	–	–	–	–	–	tr

Pl = plagioclase, **Qtz** = quartz, **Chl** = chlorite, **Ms** = muscovite, **Alm** = almandine, **Smc** = smectite, **Kln** = kaolinite, **Prh** = prehnite, **Pmp** = pumpellyite, **Cal** = calcite, **Ep** = epidote, **Hem** = hematite, **Aln** = allanite, **Py** = pyrite, **Ttn** = titanite, **Ilm** = ilemenite, **Zrn** = zircon, **Rt** = rutile, **Ap** = apatite.

+ = common **x** = considerable **o** = minor **tr** = traces **–** = missing

TABLE 2

Mean values and standard deviations of phyllosilicate "crystallinity" indices (Δ2°Θ) of the granite samples

Goniometer speed	XRD parameters	Whole rock	2μm fractions
2°/min	IC(002)	0.220 (n=1)	0.369±0.066 (n=6)
	ChC(001)	0.307 (n=1)	0.332±0.045 (n=7)
	ChC(002)	0.276±0.021 (n=11)	0.282±0.026 (n=12)
1/2°/min	IC(002)	–	0.300±0.049 (n=5)
	ChC(001)	–	0.226±0.025 (n=9)
	ChC(002)	0.229±0.012 (n=12)	0.246±0.024 (n=11)

ROCK CHEMISTRY

Chemical analyses of nine samples from the studied area are given in Table IV. It is clear that the acidic garnet-free rocks are chemically different from the garnet-rich ones. These differences are largely attributed to the abundance of the garnet and the presence of metasediment xenoliths in the latter ones. In general, these plagiogranites are characterized by high silica and very low K_2O percent. However, the silica content in samples No. 34_g, 34_{g+e}, and 35 are relatively very low resulting from an enrichment in garnet, furthermore, sample No. 36 has high K_2O percent, showing normative Or (about 20) more than normative Ab (about 18) resulting from the effect of metasediments.

In the An-Ab-Or diagram (*Fig. 3*) the analyses plot on the low pressure of feldpsar field. In the same diagram, most of samples are situated in the tonalite or trondhjemite field, except sample No. 36 which is located in the granite field, showing high K_2O percent.

The $(FO+0.9Fe_2O_3)$-(K_2O+Na_2O)-MgO diagram (AFM diagram) is used to show the differentiation trends leading to the oceanic plagiogranites. The oceanic leucocratic rocks usually follow the tholeiite trend (COLEMAN and PETERMAN 1975). The analyzed rocks show more or less the same trend, whereas sample Nos. 34_g, 34_{g+e} and 35 situate nearer to the $(FeO+0.9Fe_2O_3)$ corner showing lower total alkali and higher total iron contents (*Fig. 4*).

A plot of SiO_2% against K_2O% (*Fig. 5*) displays that most of garnet-poor granite samples fall within the oceanic plagiogranite field, except sample No. 5a which plots outside the field having lower SiO_2 and K_2O contents, that may be due to contamination with basic rocks. On the other hand, samples Nos. 34_g, 34_{g+e} and 35 plot near to the field of subalkaline oceanic basalts and gabbros due to the presence of garnet,

TABLE 3

Some phyllosilicate "crystallinity" indices measured on airdried (AD) and ethylene glycol-solvated mounts of the <2μm fractions of granite samples

goniometer	speed	2°/min						1/2°/min					
sample		1	5b	5d	34b	35	averages	1	5b	5d	34b	35	averages
IC(002)	AD	0.385	0.437	–	–	–	0.411 (n=2)	0.291	–	–	0.375	0.272	0.313 (n=3)
	EG	0.386	0.323	0.353	0.353	0.350	0.353±0.022	0.315	–	–	0.303	0.269	0.296 (n=3)
ChC(001)	AD	0.295	–	0.350	0.351	0.394	0.348±0.041	0.230	0.257	0.239	0.266	0.262	0.251±0.016 (n=5)
	EG	0.284	0.376	0.299	0.311	0.350	0.324±0.038 (n=5)	–	–	–	–	0.250	0.250 (n=1)
ChC(002)	AD	0.308	0.300	0.313	0.266	0.320	0.301±0.021 (n=5)	0.248	0.251	0.295	0.248	0.236	0.256±0.023 (n=5)
	EG	0.281	0.289	0.281	0.282	0.282	0.283±0.003 (n=5)	0.248	0.248	0.257	0.225	0.249	0.245±0.012 (n=5)

TABLE 4

Chemical analyses of the studied rocks (weight per cent).

	1	5a	5b	5d	34g	34g+e	34d	35	36
SiO_2	69.99	58.72	66.42	68.30	49.18	51.36	61.43	49.97	66.53
TiO_2	0.29	0.52	0.43	0.28	0.56	0.54	0.33	1.18	0.52
Al_2O_3	13.86	15.63	14.46	13.65	16.60	16.46	18.10	16.22	16.19
Fe_2O_3	0.49	1.08	0.61	0.60	3.67	3.23	1.69	2.25	1.52
FeO	2.63	5.50	4.29	2.90	17.80	14.02	3.37	15.68	3.96
MnO	0.05	0.19	0.11	0.05	2.25	1.65	0.71	1.48	0.22
MgO	0.72	1.44	0.89	0.80	2.35	2.21	0.08	2.94	1.63
CaO	2.22	7.32	3.31	3.78	2.52	2.74	3.36	2.56	0.31
K_2O	0.48	0.05	0.17	0.14	0.24	0.32	0.53	0.32	3.35
Na_2O	6.28	5.37	6.47	6.45	2.64	3.70	7.89	3.06	2.15
$-H_2O$	0.15	0.09	0.09	0.07	0.08	0.10	0.07	0.12	0.12
$+H_2O$	1.69	2.76	1.96	1.61	1.96	2.43	1.61	3.18	3.30
CO_2	0.03	0.83	0.39	0.79	0.37	0.37	0.42	0.04	0.00
P_2O_5	0.41	0.23	0.13	0.08	0.15	0.14	0.11	0.50	0.06
Sum	99.29	99.73	99.73	99.50	100.37	99.27	99.70	99.50	99.86
C I P W norm									
q	25.19	10.83	18.71	22.46	10.45	9.35	5.30	9.97	35.75
c	0.02	–	–	–	8.63	6.24	– 7.47	8.61	
or	2.84	0.30	1.00	0.83	1.42	1.89	3.13	1.89	19.80
ab	53.14	45.44	54.75	54.58	22.34	31.31	66.77	25.89	18.19
an	8.15	18.40	9.91	7.88	9.18	10.34	12.41	9.18	1.15
cpx	–	9.52	2.74	4.55	–	–	0.79	–	–
di	–	3.05	0.74	1.54	–	–	0.03	–	–
hd	–	6.46	2.00	3.01	–	–	0.76	–	–
opx	5.83	7.44	7.68	4.14	38.77	30.76	5.35	35.06	9.63
en	1.79	2.17	1.87	1.28	5.85	5.50	0.18	7.32	1.06
fs	4.04	5.27	5.80	2.86	32.92	25.26	5.16	27.74	5.57
mt	0.71	1.57	0.88	0.87	5.32	4.68	2.45	3.26	2.20
il	0.55	0.99	0.82	0.53	1.06	1.03	0.63	2.24	0.99
hm	–	–	–	–	–	–	–	–	–
ru	–	–	–	–	–	–	–	–	–
ap	0.97	0.54	0.31	0.19	0.36	0.33	0.26	1.18	0.14
cc	0.07	1.89	0.89	1.80	0.84	0.84	0.96	0.09	–

g = garnet-rich part,, g+e= garnet+enclave-rich part

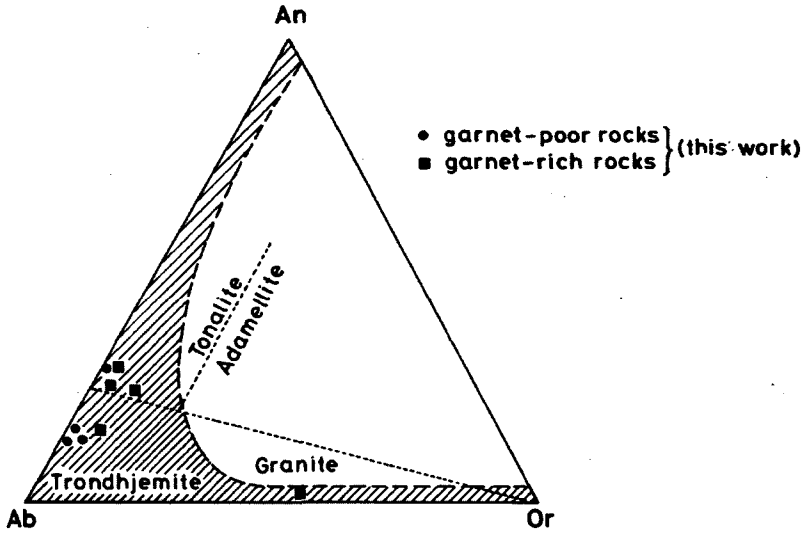

Fig. 3.
Ab-An-Or triangular diagram for plagiogranitic rocks. Shaded area: low pressure feldspar field (<5Kbar); dotted lines separate different rock types on the basis of their feldspar contents (O'CONNER 1965)

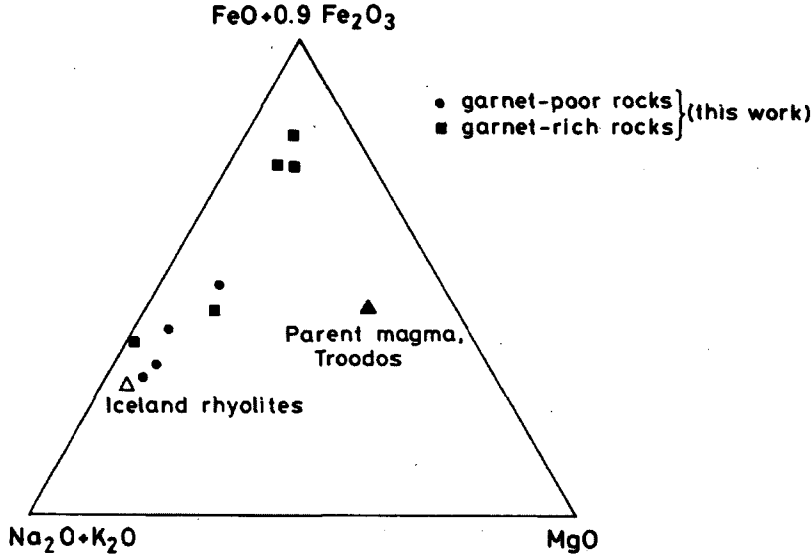

Fig. 4.
AFM diagram showing plagiogranites of the studied area comparing with the Troodos plagiogranites. Parent magma of Troodos ophiolite and Iceland rhyolites are also plotted (COLEMAN and PETERMAN 1975)

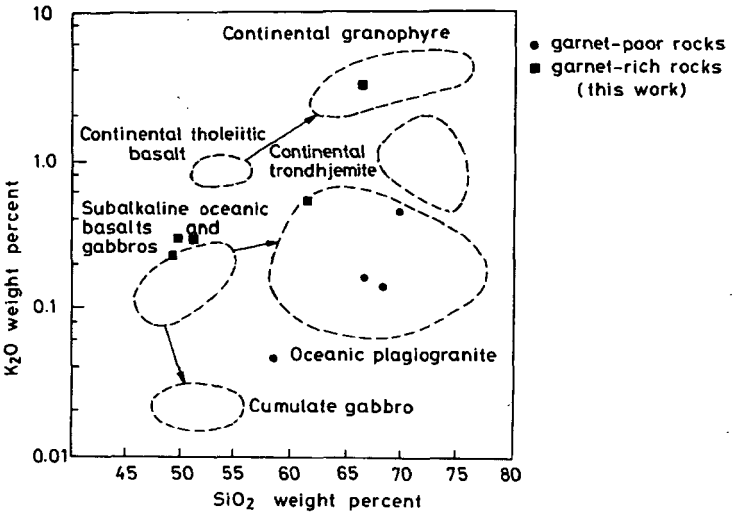

Fig. 5.
A plot of SiO_2 againts K_2O for Szarvaskő plagiogranites and common rocks of equivalent composition for comparison (COLEMAN and PETERMAN 1975).

while sample No. 36 lies within continental granophyre field as a result of metasediment contamination.

According to COLEMAN and PETERMAN (1975), the $(K_2Ox100)/(K_2O+Na_2O)$ ratio is about 49% for granophyres, but that of oceanic leucoratic rocks is about 5%. The calculated ratios for the analyzed samples mostly range from 0.9 to 9.47, so that the higher values are shown by garnet-rich rocks (in sample No. 36 this ratio reaches 60.9%).

In fact, Szarvaskő plagiogranites contain higher content of Rb (using analytical data of BALLA *et al.* 1983, DOWNES *et al.* 1990), which may be due to submarine alteration or country rock contamination. In addition, the REE patterns show significant enrichment of light REE, strong negative Eu anomalies and slight heavy REE depletion when compared with the REE patterns of Troodos oceanic plagiogranites (KAY and SENECHAL 1976).

$^{87}Sr/^{86}Sr$ ratios of Szarvaskő plagiogranites (BALLA *et al.* 1983; DOWNES *et al.* 1990) range from 0.70669–+0.00001 0.70754–+0.00001 which is obviously higher than those of Troodos (reach 0.7060; COLEMAN and PETERMAN 1975). This enrichment in isotope ratios may be attributed to rock alteration.

MINERAL CHEMISTRY

The chemistry of rock-forming minerals from plagiogranites of Szarvaskő have been determined by electron microprobe analysis (Table V).

The plagioclase of plagiogranite is characterized by high sodium and very low calcium content (average An-content vary from 1 to 4%). On the other hand, the plagioclase in metasediments that form enclaves in the granite, shows relatively higher anorthite content with an average of An_{23}. The latter exhibits notable composional

zoning with increasing Ca-content towards the grains' rims. Furthermore, the potassium content of plagioclase is very low.

Chlorite usually occurs as replacement of original biotite or as filling in veins and cracks of plagioclases. Chlorite of metasediment xenoliths is very similar to those occurring in acidic rocks. The only detectable difference is an increasing in potassium-content (average K of 0.247) in the former case, comparing with the latter case, where the average K varies from 0.011 to 0.045.

Prehnite generally exists an infilling veins and cracks and also as alteration products of both biotite and plagioclase. Two types of prehnites could be observed, the first is Fe-rich with average Fe_2O_3 content of 4.65, whereas the second type is Fe-poor with average Fe_2O_3 content of 0.51.

Epidote analyses refer to considerable amounts of REE. In fact, both epidotes and allanites were formed after biotite and plagioclase, respectively.

The analyzed garnets from garnet-poor granites and from garnet-rich rocks exhibit the same chemical composition. They are almandine-rich, with composition ranges of Alm_{75-76} Pyr_{7-11} $Spess_{9-14}$ $Gross_{4-5}$. Garnets frequently show slight, irregular chemical zoning without significant change in composition from core to rim. The distribution of Mg, Fe, Mn and Ca in the individual zoned crystals from both rock types are shown in *Figs. 6* and *7*. From *Fig 6.* it is obvious that Fe is the only element displaying an increase towards the rim, while Mg, Mn and Ca vary in irregular patterns. In *Fig. 7a*, both Fe and Mn increase towards the rims, coupled with a decrease in Mg content, while Ca exhibits slight change. In an other grain (*Fig. 7b*) Fe changes against Ca, while both Mg and Mn do not show significant difference between the core and rim. In the Alm-Pyr-(Spess+Gross) triangular diagram, the garnets of this work are located within the field of metapelites (*Fig. 9*). Moreover, using the $(Mg^{++}+Fe^{++})$–Mn^{++}–Ca^{++} triangular diagram (*Fig. 8*), these garnets fall between the fields of igneous silicic rocks and the representive points of the garnet from the metamorphic rocks which surround the Vedrette di Ries pluton (BELLINI *et al.* 1979).

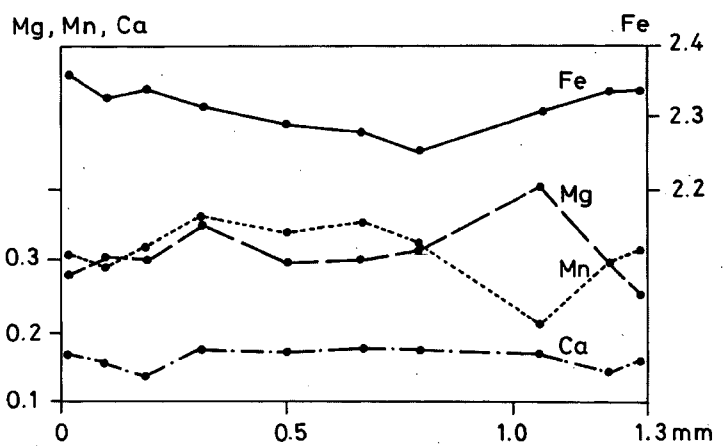

Fig. 6.
Zoning of a garnet grain from garnet-poor granites of Szarvaskő.

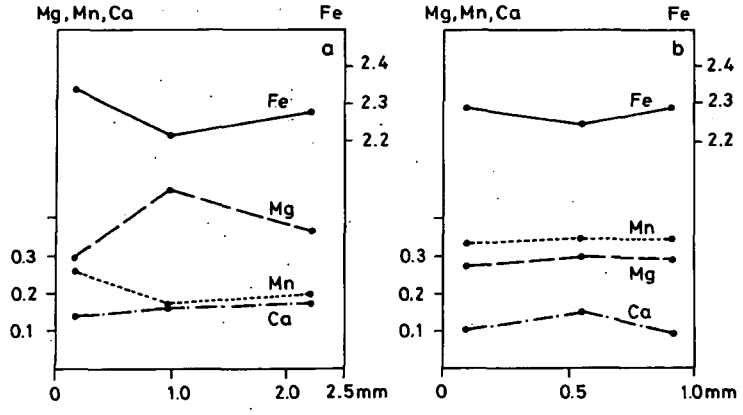

Fig. 7 (a–b)
Zoning of garnet grains from garnet-rich granites of Szarvaskő

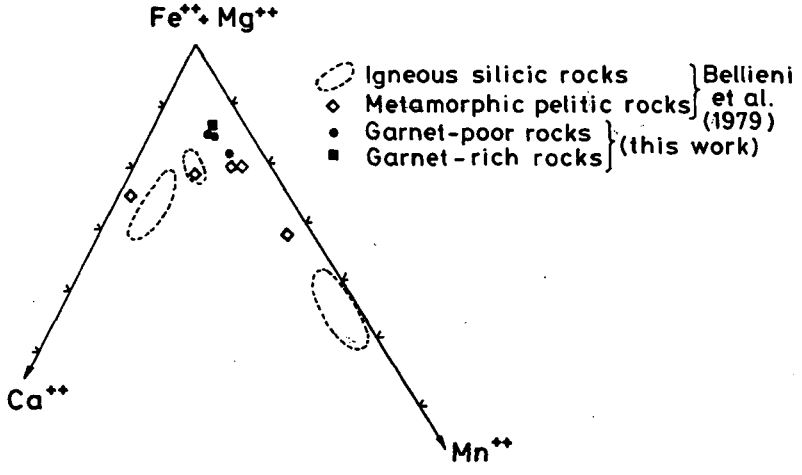

Fig. 8.
A plot of $(Fe^{++}+Mg^{++})$–Mn^{++}–Ca^{++} (after BELLIENI *et al.* 1979) illustrating the composition of the studied garnet

TABLE 5

Electron microprobe data of rock minerals in weight percent and number of cations per unit cell plagioclase analyses

sample	1	5b	5d	34b	34e	34d
n	4	4	4	4	3	4
SiO_2	67.61±0.96	67.21±0.26	69.16±2.13	66.66±1.47	62.31–60.48: 65.68**	68.23±0.33
Al_2O_3	20.46±0.85	20.25±0.12	19.16±1.48	20.57±0.49	24.05–21.91: 25.59	20.10±0.21
CaO	00.60±0.64	0.32±0.11	0.15±0.10	1.10±0.30	4.77– 2.04: 6.32	0.11±0.01
Na_2O	11.38±0.49	11.03±0.64	11.33±0.70	10.58±0.67	8.79– 7.92: 10.34	2.20±0.27
K_2O	0.11±0.04	0.56±0.58	0.09±0.06	0.23±0.14	0.16– 0.13: 0.22	0.06±0.02
Total	99.68±0.26	99.36±0.28	100.15±0.24	99.13±1.92	100.08–99.69:100.62	100.69±0.50
			numbers of cations on the basis of 8 (0)			
Si	2.948±0.039	2.959±0.004	3.008±0.077	2.196±0.039	2.753–2.884:2.672**	2.965±0.011
Al	1.051±0.041	1.051±0.009	0.996±0.080	1.069±0.014	1.257–1.332:1.121	1.029±0.011
Ca	0.029±0.030	0.015±0.005	0.007±0.005	0.037±0.023	0.226–0.299:0.096	0.005±0.002
Na	0.962±0.043	0.942±0.053	0.956±0.062	0.905±0.058	0.752–0.880:0.682	1.028±0.020
K	0.006±0.002	0.031±0.033	0.005±0.003	0.013±0.008	0.009–0.012:0.008	0.003±0.001
Total	4.996±0.011	4.999±0.015	4.971±0.068	4.939±0.111	4.992–4.980:5.004	5.031±0.016
	An_3	An_2	An_1	An_4	An_{23}	An_1

e = enclave metasediment n = number of analyses ** the highest and the lowest values

Garnet analyses

	1	5a	5b	34b
n	6	5	10	6
SiO_2	36.18±0.29	37.04±0.28	36.04±0.33	37.03±0.17
Al_2O_3	21.04±0.17	20.80±0.13	20.97±0.20	20.80±0.23
FeO	32.50±1.91	33.41±0.34	33.91±0.52	33.65±0.56
MgO	1.75±0.50	2.26±0.09	2.62±0.29	2.72±0.70
MnO	5.95±3.17	4.41±0.38	4.22±0.63	3.97±1.10
CaO	1.46±0.37	1.79±0.05	1.86±0.16	1.56±0.42
Total	98.88±1.85	99.71±0.05	99.61±0.31	99.73±0.38
	numbers of cations			
Si	2.975±1.014	3.007±0.008	2.942±0.020	2.995±0.017
Al	2.039±0.010	1.990±0.009	2.018±0.018	1.982±0.015
$*Fe^{2+}$	2.235±0.121	2.268±0.021	2.315±0.032	2.276±0.041
Mg	0.214±0.016	0.273±0.010	0.318±0.034	0.329±0.080
Mn	0.415±0.221	0.303±0.026	0.292±0.045	0.272±0.075
Ca	0.128±0.033	0.156±0.004	0.163±0.014	0.135±0.035
Total	8.006±0.015	7.998±0.006	8.048±0.013	7.988±0.011
	Pyr_7	Pyr_9	Pyr_{10}	Pyr_{11}
	Alm_{76}	Alm_{76}	Alm_{75}	Alm_{76}
	$Spess_{14}$	$Spess_{10}$	$Spess_9$	$Spess_9$
	$Gross_4$	$Gross_5$	$Gross_5$	$Gross_4$

* total iron calculated as FeO and Fe^{2+}, respectively n = number of analyses

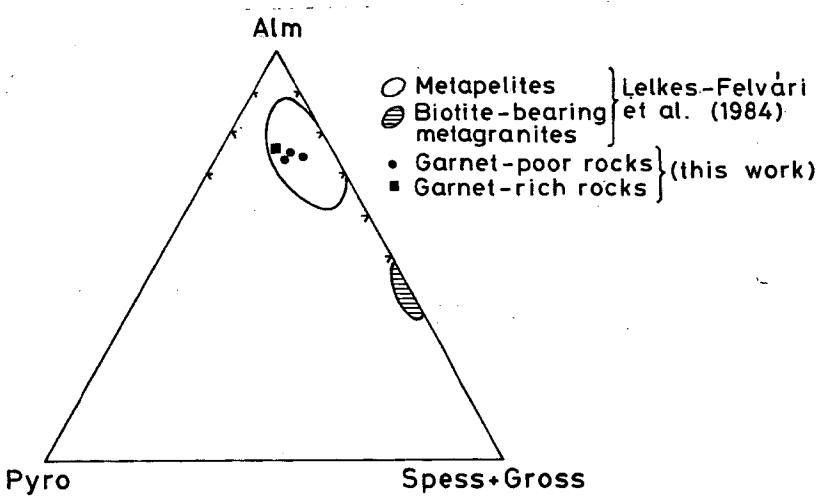

Fig. 9.
Alm-Pyr-(Spess+Gross) diagram of LELKES-FELVÁRI *et al.* (1984) showing the distribution of the studied garnet

TABLE 5. contd

Chlorite analyses

sample	1	5b	5d	34b	34e	34d
n	9	7	4	4	2	4
SiO_2	27.21±0.72	24.45±0.88	26.33±0.83	26.20±0.68	26.94–26.81:27.07**	25.69±1.45
Al_2O_3	17.64±0.72	19.83±1.08	17.60±0.60	17.91±0.44	17.89–17.67:18.11	18.79±1.69
*FeO	31.54±2.24	36.18±1.41	31.51±0.27	36.06±2.36	33.13–32.80:33.45	34.94±1.20
MgO	8.94±1.91	7.02±0.67	8.56±0.34	6.73±1.09	6.48– 6.46: 6.50	6.72±1.69
MnO	0.69±0.34	0.38±0.14	0.55±0.12	0.50±0.02	0.49– 0.48: 0.49	0.42±0.07
CaO	0.750.34	0.24±0.19	0.67±0.14	0.52±0.43	0.72– 0.51:0.93	0.94±1.18
Na_2O	0.04±0.01	0.03±0.01	0.04±0.04	0.05±0.04	0.04– 0.03: 0.05	0.05±0.04
K_2O	0.16±0.14	0.04±0.03	0.12±0.07	0.14±0.09	0.88– 0.55: 1.21	0.07±0.03
TiO_2	1.54±0.57	0.27±0.34	1.02±0.26	0.95±0.90	1.59– 1.53: 1.65	0.49±0.62
Total	88.52±0.87	88.45±0.64	86.41±0.71	89.05±0.45	88.16–87.41:88.89	88.11±0.91
			numbers of cations on the basis of 28 (0)			
Si	5.898±0.129	5.445±0.178	5.864±0.128	5.786±0.078	5.938– 5.927: 5.948**	5.704±0.235
Al^{IV}	2.102±0.129	2.555±0.178	2.136±0.128	2.214±0.078	2.062– 2.073: 2.052	2.296±0.235
Al^{VI}	2.406±0.092	2.651±0.115	2.481±0.109	2.448±0.086	2.586– 2.569: 2.602	2.626±0.293
*Fe	5.722±0.446	6.740±0.289	5.869±0.096	6.667±0.550	6.106– 6.086: 6.126	6.498±0.487
Mg	2.887±0.590	2.330±0.213	2.844±0.124	2.211±0.355	2.129– 2.122: 2.135	2.226±0.557
Mn	0.127±0.600	0.073±0.026	0.103±0.022	0.094±0.004	0.090– 0.089: 0.091	0.080±0.014
Ti	0.251±0.094	0.046±0.057	0.171±0.044	0.156±0.146	0.264– 0.252: 0.276	0.082±0.104
Ca	0.137±0.078	0.056±0.046	0.161±0.032	0.121±0.100	0.170– 0.119: 0.220	0.221±0.273
Na	0.018±0.005	0.014±0.006	0.019±0.018	0.021±0.015	0.018– 0.014: 0.022	0.023±0.016
K	0.045±0.038	0.011±0.009	0.035±0.019	0.040±0.023	0.247– 0.156: 0.337	0.021±0.009
Total	19.621±0.125	19920±0.095	19.684±0.077	19.758±0.125	19.608–19.555:19.661	19.776±0.157

e = enclave metasediment n = number of analyses * total iron calculated as FeO and Fe^{2+}, respectively ** the highest and the lowest values

Prehnite analyses

sample	1	5b dark	5d light	5d	34b	34d
n	4	4	4	4	3	5
SiO_2	43.91±0.17	43.47±0.24	42.69±0.32	43.10±0.50	44.49–44.00:44.84**	42.81±0.54
Al_2O_3	23.39±0.96	24.15±0.17	21.21±0.79	23.74±0.11	23.64–23.44:23.77	22.65±1.48
*Fe_2O_3	0.94±0.62	0.51±0.10	4.65±0.90	1.44±0.73	0.25– 0.21: 0.31	1.73±1.74
MgO	0.04±0.08	0.01±0.01	0.01±0.01	0.10±0.05	0.00– 0.00	0.01±0.02
MnO	0.03±0.01	0.02±0.02	0.00±0.00	0.01±0.01	0.02– 0.00: 0.04	0.03±0.02
CaO	26.97±0.16	27.61±0.22	27.11±0.24	27.39±0.24	26.85–26.20:27.26	27.26±0.70
Na_2O	0.13±0.11	0.04±0.10	0.03±0.01	0.02±0.01	0.35– 0.10: 0.84	0.06±0.07
K_2O	0.02±0.01	0.01±0.01	0.01±0.01	0.03±0.01	0.04– 0.02: 0.05	0.03±0.05
TiO_2	0.05±0.00	0.07±0.05	0.04±0.03	0.06±0.02	0.02– 0.00: 0.03	0.02±0.03
Total	95.49±1.11	95.89±0.41	95.75±0.48	95.88±1.09	95.66–95.03:96.20	95.59±1.37
			numbers of cations on the basis of 22 (O)			
Si	6.059±0.052	5.977±0.011	5.970±0.006	5.947±0.024	6.107– 6.087: 6.118**	6.000±0.019
Al	3.803±0.117	3.914±0.009	3.498±0.120	3.861±0.044	3.826– 3.815: 3.840	3.740±0.213
*Fe^{3+}	0.098±0.065	0.053±0.011	0.496±0.099	0.150±0.074	0.026– 0.022: 0.032	0.183±0.187
Mg	0.009±0.016	0.002±0.002	0.003±0.003	0.021±0.011	0.000– 0.000	0.002±0.001
Mn	0.004±0.001	0.002±0.002	0.001±0.001	0.001±0.001	0.002– 0.000: 0.005	0.002±0.001
Ca	3.989±0.029	4.70±0.023	4.064±0.008	4.051±0.061	3.951– 3.849: 4.018	4.095±0.065
Na	0.035±0.030	0.011±0.002	0.008±0.002	0.006±0.004	0.092– 0.026: 0.223	0.017±0.018
K	0.004±0.001	0.002±0.001	0.003±0.001	0.006±0.002	0.006– 0.004: 0.009	0.006±0.009
Ti	0.006±0.011	0.007±0.005	0.004±0.003	0.005±0.003	0.002– 0.000: 0.003	0.002±0.003
Total	14.005±0.020	14.038±0.010	14.038±0.005	14.048±0.020	14.013–13.976:14.065	14.049±0.025

* total iron calculted as Fe_2O_3 and Fe^{3+}, respectively ** the highest and the lowest values **n** = number of analyses

TABLE 5. contd.

Epidote analyses

sample	1	34b	34d
n	2	2	3
SiO_2	37.41–37.19:37.63**	36.36–38.03:38.69**	37.87–37.33:38.42**
Al_2O_3	24.50–24.35:24.65	24.58–24.45:24.71	23.98–23.46:24.32
*Fe_2O_3	11.72–11.62:11.80	12.19–11.83:12.54	10.04– 5.81:12.22
MgO	0.05– 0.03: 0.07	0.03– 0.00: 0.06	0.80– 2.20: 0.04
MnO	0.10– 0.06: 0.13	0.31– 0.27: 0.34	0.24– 0.05: 0.22
CaO	23.32–23.03:23.59	23.47–23.19:23.75	23.08–22.82:23.36
Na_2O	0.04– 0.04: 0.04	0.06– 0.04: 0.07	0.01– 0.00: 0.02
K_2O	0.00– 0.00	0.03– 0.02: 0.04	0.00– 0.00
TiO_2	0.21– 0.19: 0.22	0.30– 0.23: 0.36	0.61– 0.00: 1.43
Total	97.33–97.13:97.52	99.32–99.09:99.55	96.54–92.64:98.55
	numbers of cations on the basis of 25 (O)		
Si	5.962– 6.938: 8.985**	5.997– 5.967: 6.026**	6.052– 6.140: 5.979**
Al	4.602– 4.565: 4.639	4.529– 4.522: 4.536	4.518– 4.368 4.684
*Fe^{3+}	1.408– 1.395: 1.420	1.436– 1.388: 1.483	1.201– 0.719: 1.446
Mg	0.012– 0.007: 0.017	0.007– 0.000: 0.014	0.194– 0.009: 0.539
Mn	0.013– 0.008: 0.018	0.041– 0.036: 0.045	0.021– 0.007: 0.029
Ca	3.982– 3.942: 4.022	3.933– 3.900: 3.965	3.955– 3.843: 4.066
Na	0.012– 0.012: 0.012	0.017– 0.012: 0.021	0.003– 0.000: 0.006
K	0.000– 0.000	0.006– 0.004: 0.008	0.000– 0.000
Ti	0.025– 0.023: 0.026	0.035– 0.027: 0.042	0.072– 0.000: 0.170
Total	16.015–16.012:16.017	15.998–15.994:16.002	16.017–15.942:16.161

* total iron calculted as Fe_2O_3 and Fe^{3+}, respectively ** the highest and the lowest values
n = number of analyses

TABLE 5. contd.

Allanite analyses

sample	1	5d
n	3	3
SiO_2	33.69– 32.98: 34.36**	34.71–34.50: 34.72**
Al_2O_3	22.51– 22.17: 22.82	23.01–22.76: 23.83
*Fe_2O_3	11.05– 10.23: 11.92	11.52–11.31: 11.94
MgO	0.35– 0.08: 11.92	0.11– 0.08: 0.12
CaO	14.53– 13.89: 15.75	15.53–14.80: 16.62
MnO	0.00– 0.00	0.34– 0.27: 0.38
TiO_2	0.58– 0.20: 1.47	0.00– 0.00
Y_2O_3	0.56– 0.31: 1.02	4.17– 4.40: 4.61
La_2O_3	2.96– 2.15: 3.39	1.31– 0.85: 1.54
Ce_2O_3	9.10– 6.44: 1.64	4.30– 2.68: 5.18
Pr_2O_3	1.00– 0.80: 1.25	0.41– 0.34: 0.49
Nd_2O_3	3.64– 3.22: 3.82	1.94– 1.33: 2.33
Sn_2O_3	0.78– 0.82: 0.70	0.68– 0.45: 0.92
Eu_2O_3	0.08– 0.08: 0.09	0.02– 0.00: 0.05
Gd_2O_3	0.42– 0.41: 0.44	0.50– 0.46: 0.58
Dy_2O_3	0.19– 0.12: 0.28	0.72– 0.65: 0.80
Ar_{23}	0.00– 0.00	0.37– 0.28: 0.43
Yb_2O_3	0.00– 0.00	0.37– 0.24: 0.51
Total	101.58–100.31:103.08	100.12–98.71:101.11
numbers of cations on the basis of 25 (O)		
Si	5.731– 5.708: 5.774**	5.816– 5.783: 5.834**
Al	4.515– 4.466: 4.553	4.545– 4.483: 4.631
*Fe^{3+}	1.427– 1.296: 1.520	1.455– 1.428: 1.503
Mg	0.088– 0.021: 0.181	0.027– 0.021: 0.030
Ca	2.646– 2.518: 2.835	2.788– 2.665: 2.954
Mn	0.000– 0.000	0.048– 0.038: 0.053
Ti	0.074— 0.009: 0.186	0.000– 0.000
Y	0.050– 0.028: 0.091	0.371– 0.310: 0.411
La	0.186– 0.133: 0.217	0.080– 0.052: 0.095
Ca	0.567– 0.396: 1.674	0.263– 0.164: 0.317
Pr	0.062– 0.049: 0.079	0.025– 0.021: 0.030
Nd	0.221– 0.193: 0.237	0.116– 0.080: 0.139
Sm	0.046– 0.041: 0.50	0.039– 0.026: 0.053
Eu	0.005– 0.005: 0.005	0.001– 0.000: 0.003
Gd	0.023– 0.023: 0.024	0.028– 0.025: 0.032
– Dy	0.011– 0.007: 0.016	0.039– 0.035: 0.043
Yb	0.000– 0.000	0.019– 0.015: 0.022
Total	15.680–15.636:15.751	15.644–15.622:15.685

n = number of analyses
* total iron calculted as Fe_2O_3 and Fe^{3+}, respectively
** the highest and the lowest values

DISCUSSION

1. The primary nature of plagiogranites

The plagiogranites of the Szarvaskő area are usually occur in small amounts as nests within the gabbroic rocks of the ophiolitic sequence, showing sharp contact with them.

Mineralogically, the investigated rocks are made of quartz and unzoned albitized plagioclase, with minor biotite which were completely replaced by secondary minerals. In comparison with the oceanic plagiogranites of Troodos (COLEMAN and PETERMAN 1975), the Szarvaskő plagiogranites differ in the presence of unzoned plagioclase with albitic compositions, and in the lack of pyroxene and amphibole (the primary ferromagnesian minerals are represented only by biotite). BAILEY and BLAKE (1974) described zoned plagioclase phynocrysts (An_{50-62}) in a matrix of albite and quartz in the extrusive silicic rocks, while in the intrusive ones referred to the presence of albitized plagioclase phenocrysts. This may be explained by the post magmatic replacement of early formed calcic plagioclase. Moreover, the above authors notes small amounts of biotite in addition to amphibole in coarser varieties (see also GILLULY 1933).

The major element abundances in Szarvaskő plagiogranites is similar to those of oceanic ones showing high silica, moderate alumina, low total Fe+Mg, extremely low K_2O contents. When comparing with the data of COLEMAN and PETERMAN (1975), normative Or is usually less than 4 except sample No. 36 which contains metasediment xenoliths showing high normative Or (~20%), whereas normative An is too small, rarely reaching 18% (An_{21-61} in Troodos plagiogranites), that may be resulted from albitization of early formed plagioclase. In general, Szarvaskő rocks usually occupy the oceanic plagiogranitic field (*Figs. 3* and *5*).

The Rb content in the studied rocks is relatively high (reaches 15 pmm). Comparing with other oceanic silicic rocks the enrichment of Rb may be due to submarine alteration or due to the contamination by country rocks (DOWNES *et al.* 1990). In the quartz-monzonitic dikelet of the Central Indian ridge, the concentration of Rb is 84 ppm (ENGEL and FISHER 1975). Significant light REE enrichment of Szarvaskő plagiogranite compared with Troodos plagiogranite may be attributed to country rock contamination. Higher values of $^{87}Sr/^{86}Sr$ may be resulted from alteration as shown by COLEMAN and PETERMAN (1975).

From the above discussion it can be suggested that the Szarvaskő plagiogranites are primary oceanic plagiogranitic rocks crystallized from late differentiated basic melt.

2. The origin of the garnet

The composition of the garnet from garnet-poor rocks is very similar to that of garnet-rich varieties: both being almandine-rich. They occur as hypidiomorphic crystals with slightly resorbed outlines and show varying zoning patterns. In general, Fe content increases and Ca content decreases towards the rims. The inclusions are represented by quartz, ilmenite, sphene and rutile.

In garnet-rich varieties, garnets usually accumulated in the granitic rocks near or at the contact with the metasedimentary enclaves. On the other hand, however garnets do not form contact metamorphic rims in metasedimentary enclaves. They are always dispersed in granitic melt.

According to their chemistry, the investigated garnets are located either within the metapelite field or near to the igneous silicic rock field (*Fig. 8* and *9*).

From the above observations it can be concluded that garnets formed by local assimilation of the country rocks. This conclusion supports the former opinion of BALLA *et al.* (1983), but contrasts with the metasomatic origin suggested by EMBEY-ISZTIN *et al.* (1985).

3. Post-magmatic alteration of plagiogranite

Most of the primary magmatic minerals of the plagiogranites are obviously altered. Thus, the original composition of plagioclase which crystallized from magmatic melt is not known, as it was completely albitized, presumably by post-magmatic processes. In addition, prehnite and white mica (sericite) were formed, mainly along cleavage surfaces and cracks of plagioclase. Biotite totally altered to chlorite, epidote, prehnite and titanite. It has been shown that the Szarvaskő plagiogranites are characterized by secondary mineral associations commonly including chlorite, epidote, prehnite, albite, calcite, titanite and sericite, while pumpellyite is completely lacking.

As proved by ÁRKAI (1983), the metabasic rocks associated with these plagiogranites were affected by pumpellyite-prehnite-quartz facies metamorphism. The averages of IC(002), ChC(001) and ChC(002) measured on $<2\mu$m fractions, using a goniometer speed of 2°/min, are 0.369±0.066, 0.329±0.042 and 0.282±0.026 Δ°28, respectively. IC(002) average value indicates anchizonal regional metamorphism (see the zone boundaries of KÜBLER 1975) or falls into the zone of deep diagenesis (see ÁRKAI, 1987). On the basis of IC, the temperature must be of ca. 230°C. Moreover, using the ChC(001) and ChC(002) anchizonal boundaries given by ÁRKAI (1991), the chlorite crystallinity of the first and second basal reflections of the studied rocks suggest anchizonal regional metamorphism of temperature of about 280°C which corresponds to the conditions of the pumpellyite-prehnite-quartz facies at low geothermal gradient or to the prehnite-actinolite facies at high geothermal metamorphic gradient (very low pressure facies). Using the chlorite geothermometer of CATHELINEAU (1988) elaborated originally for andesitic rocks and proved later on for metabasic rocks and metasandstones, the calculated temperature values of the metamorphism of the Szarvaskő plagiogranites vary between 276 and 350°C (in average, 298°C). The variations in chemical compositions of certain secondary minerals (prehnite, chlorite) may refer to local changes in equilibrium conditions.

Finally, it may be concluded that the secondary mineral assemblage of the plagiogranite might form in three various metamorphic regimes, i. e., during ocean-floor hydrothermal, burial and regional (dynamothermal) processes, from which the last Cretaceous regional metamorphic event was already proved (ÁRKAI 1983). The superposition of these processes seems to be also possible. Local metasomatic effects resulting from the circulating hydrothermal solutions and/or from the surrounding sedimentary sequence and/or from sea water interactions might also contribute to the formation of secondary minerals. The microstructural features and the mineral assemblage of prehnite + epidote + chlorite + albite + quartz of the plagiogranite alone do not give sufficient evidence either for determining exactly the metamorphic facies or the paleotectonic system of the transformation. Chlorite "crystallinity" values together with the prehnite + epidote + chlorite + albite + quartz assemblage may suggest metamorphic condition corresponding to the low pressure type prehnite-actinolite facies of LIOU *et al.* (1985) who mentioned that other assemblages characteristic of this facies include epidote + prehnite ± hematite ± magnetite could be recognized in active geothermal systems.

CONCLUSIONS

Based on new petrographic, bulk chemical and mineral chemical data the following conclusions can be drawn on the petrogenesis of the granitic rocks associated with basic-ultrabasic sequence of the Jurassic Szarvaskő complex, Bükk Mountains, Hungary.

1. Major element bulk chemical compositions of the investigated granitic rocks is similar to that of the oceanic plagiogranites, and refer to crystallization from late differentiated basic melt.

2. Almandine-rich garnet found in varying proportions in granite is accumulated near to or at the contacts with metasedimentary (biotite-plagioclase hornfels) enclaves. As no direct spatial connection can be observed between the garnets and the enclaves (garnet are always dispersed in the granitic melt), the formation of garnet is explained by local assimilation of the country rocks in a deeper level and not by direct (contact metamorphic or metasomatic) effects caused by the granitic melts on the enclaves.

3. The primary (magmatic) minerals of the plagiogranite was commonly altered: plagioclase was completely albitised (±prehnite, white mica), biotite was totally altered to chlorite, epidote, prehnite and titanite. The secondary mineral assemblage (quartz, albite, chlorite, epidote, prehnite, white mica, calcite and titanite), illite and chlorite "crystallinity" indices as well as the data obtained by the chlorite-Al" geothermometer refer to low temperature (ca. 280°–300°C) metamorphism. Further studies on the whole ophiolite-like sequence and the surrounding sediments are needed to prove eventual ocean-floor (hydrothermal) or burial metamorphic events which might precede the Cretaceous regional (dynamothermal) effect documented by ÁRKAI (1983).

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TERTIARY GRANITIC ROCKS ALONG THE SOUTHERN MARGIN OF THE PANNONIAN BASIN

V. KNEŽEVIĆ,* S. KARAMATA,* V. CVETKOVIĆ*

Faculty of Mining and Geology, University of Belgrade

ABSTRACT

The Oligocene-Miocene granitic magmatism of the Bukulja-Cer-Motajica (BCM) belt developed as follows:

– from about 30 Ma until 25 Ma ago I-type magmas gave in the BCM belt the hornblende bearing granodiorites of Brajkovac and Bukulja, the tonalites of Bogatic and the deep seated facies of the quartz monzonites of Cer as the first and the biotite granites of Bukulja and the dyke rocks of Brajkovac as the final products.

– around 20 Ma ago strng heating along the BCM belt induced mobilization to melting of older granitic rocks. The reomorphic intrusion of mobilized/initially melted rock masses gave the quartz monzonite of Cer and from the intruded highly melted masses originated the biotite graniodiorite of Stražanica.

– about 18 Ma ago S-type magmas were intruded and gave muscovite bearing granites of Bukulja, Cer and Brajkovac (?).

– the magmatism of the first two phases is widespread in all the Central Balkan peninsula, the youngest S-granitic phase is restricted to the BCM belt only because of specific geotectonic conditions along the southern margin of the Pannonian basin.

INTRODUCTION

At the south of the river Sava (and Danube) parallel to the southern margin of the Pannonian basin is situated a belt of Tertiary granitic rocks, extending from Bukulja at the east, westwards to Brajkovac, Cer, Stražanica, Bogatić, all in Serbia, and further to Motajica ad Prosara in northern Bosnia. These granites differ from other Tertiary granitic rocks of the Serbian-macedonian massif, the Vardar zone and the Dinarides in compositions, age, originand geotectonic affinities and in metallogeny. This was the reason to study them in details.

Following the field work, mineralogical and chemical analyses performed by the autjors, the geochemistry and isotope ages were studied. The trace alaments have been determined by Mrs. TATJANA MARCHENKO, IGEM of the Russian Acad. of Sciences, Moscow, and by the spectrochemical laboratory of the GEOINSTITUTE, Belgrade, and the isotopic ages and Sr-isotope ratios by Dr. Z PÉCSKAY, ATOMKI, Hungarian Acad. of Sciences, Debrecen, and dr. YU PUSHKAREV, VSEGEI, Sankt Petersburg. The Serbian Academy of Sciences and Arts, Belgrade enabled the work by the grant "Magmatic rocks of Serbia" as a part of the project "Geodynamics". The authors express their gratitude to the above mentioned colleagues and institutions.

* YU–11000 Belgrade, Dušina 7.

The granitic rocks of the Oligocene-Miocene Bukulja–Cer–Motajica (BCM) belt are exposed in Serbia at the surface as the massifs and bodies of Bukulja, Brajkovac, Cer and Stražanica and were detected by drillholes and geophysics near Bogatić (*Fig. 1*). They occur further westwards in northern Bosnia as the massif of Motajica and small bodies at Prosara. All these rocks exhibit some common characteristics, but differences are also evident. The granitic rocks of northern Bosnia are less studied and need further investigation, for this reason they will not be presented here.

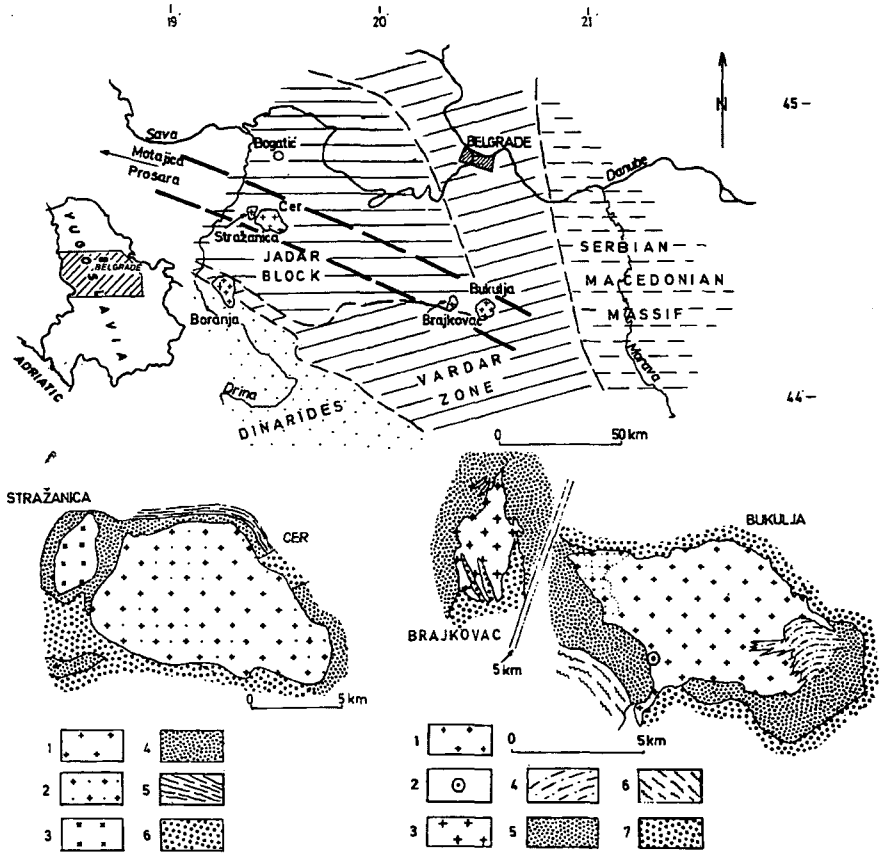

Fig. 1

The position of the Bukulja–Cer Motajica belt of Tertiary granitic rocks. At the lower part of the figure are presented the sketches of the Bukulja and Brajkovac granitic massifs (right) and of the Cer and Stražanica granitic massifs (left).

Legend for the lower left sketch: 1 – quartz monzonite; 2 – leucocratic and muscovite granites; 3 – granodiorite; 4 – conact metamorphic rocks; 5. – Paleozoic; 6 – Neogene and Quarternary. Legend for the lower right sketch: 1 – two mica granite with biotite granite in the deep creeks, leucocratic facies pointed; 2 – drillholes with granodiorite enclaves; 3 – granodiorite of Stražanica; 4 – feldspatizied schists at the contact; 5 – contact metamorphic rocks; 6 – Devonian – Carboniferous; 7 – Neogene and Quarternary.

The BCM belt crosscuts the geotectonic framework built in Upper Cretaceous or even at early Paleogene, i. e. the belt was formed along a young Oligocene-Miocene fracture zone.

The granites are intrusive into Paleozoic, Triassic and Cretaceous sedimentary rocks, exhibiting at them contact metamorphism of different intensity. Locally (at Brajkovac and Bogatić) they are associated with Oligocene andesites and dacites, but the younger Miocene intrusives are without volcanic analogues.

The granitic bodies were exposed at the surface about 12 Ma ago, and the rocks from their contact aureoles and all existing granitic varieties are deposited in Uppermost Middle Miocene sedimentary formations.

PETROLOGY AND GEOCHEMISTRY

The granitic rocks show a wide span of varieties: quartz diorites, tonalites and granodiorites as well as biotite, muscovite and leucocratic granites, sometimes even in the same body (*Fig. 1* and *2*).

A short description of the individual granitic bodies follows, but to avoid descriptions and to shorten the text the mineralogic composition and main chemical data of the varieties in the studied granitic bodies are presented at the Table 1, and the isotopic age data and isotopic Sr-ratios at the Rable 2 (all according to our investigations and the data from KENŽEVIĆ 1962, KOSTIĆ and PAVLOVIĆ 1978, STEIGER *et al.* 1989; KARAMATA *et al.* 1990, DIVLJAN and CVETIĆ 1991; VUKOV and MILIVOJEVIĆ 1993 as well

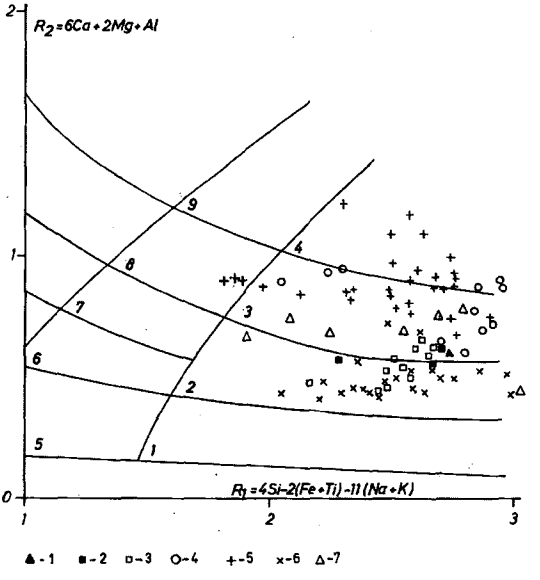

Fig. 2.

Distribution of rock varieties in the granitic massifs of the Bukulja–Cer–Motajica belt in the R_1–R_2 diagram (according to LA ROCHE *et al.* 1980).

Fields: 1– alkali granite; 2 – granite; 3 – granodiorite; 4 – tonalite; 5 – quartz alkali syenite; 6 – quartz syenite; 7 – quartz monzonite; 8 – quartz monzodiorite; 9 – quartz diorite; Signs: 1 – Bukulja granodiorite; 2 – biotite granite of Bukulja; 3 – two mica granite of Bukulja; 4 – granodiorite of Brajkovac; 5 – quartz monzonite of Cer; 6 – leucocratic and muscovite granites of Cer; 7 – granodiorite of Stražanica.

as unpublished data by Z. PÉCSKAY and YU. PUSHKAREV). At the table 1 is given also the intensity of the contact metamorphism associated to each rock type. The same data for the Oligocene granodioritic massif of Boranja (after KARAMATA 1955, KARAMATA *et al.* 1992), as a typical example for the Tertiary granitic rocks in Dinarides originated in one phase are presented for comparison at the Tables 1 and 2.

The granitic massif of Bukulja (*Fig. 1*) consists mainly of two mica and muscovite granites, but in deep creeks biotite granites occur too. It is important that hornblende biotite granodiorites and quartz diorites were found in some drillholes as enclaves (?) within the granites. The two mica granites exhibit strong contact metamorphism and feldspatization of surrounding rocks. The granodiorites and biotite granites correspond to I-type, but the muscovite bearing granites to the S-type (Tables 1 and 2). The I-typerocks show a magmatic (volcanic) arc affinity, the S-type ones have character of "collisional granites" (*Fig. 3*). The biotite from the biotite granite shows a K/Ar age 27–25 Ma, the whole rock analyses give however K/Ar age of only 18 Ma probably because of a K-feldspar rejuvenation. In the two-mica granites the biotite, muscovite as well as the K-feldspar give 19–15 Ma. The data on biotite from the biotite granite are probably close to the age of their intrusion possibly only somewhat younger. The biotite granitic magmas represent probably the last most acidic differentiates of magmas which gave the hornblende bearing rocks. The ages of the minerals from the two-mica granites (muscovite bearing granites of S-type) correspond to the crystalliza-

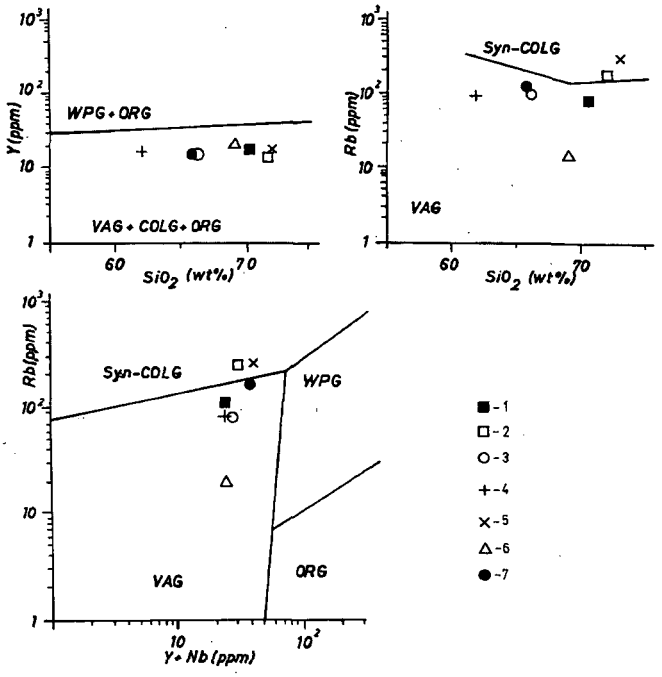

Fig. 3.
Distribution of data (mean values) for the granitic rocks of the Bukulja–Cer–Motajica belt in the discrimination diagrams for geotectonic affinity (after PEARCE et al. 1984).
Signs: 1 – biotite granite of Bukulja (10); 2 – two mica granite of Bukulja (17); 3 – granodiorite of Brajkovac (26); 4 – quartz monzonite of Cer (36); 5 – leucocratic and muscovite granites of Cer (43); 6 – granodiorite of Stražanica (4); 7 – granodiorite of Boranja (81).

TABLE 1

Mineralogy, main chemical data and data on contact metamorphism

	Contact metamorphism	Q	Kf*	Pl	Ho	Bi	Mu	Tn	Ort	Ap	Zr	MM	Ep	Tour	Gr	SiO_2	CaO	Na_2O/K_2O
BUKULJA,																		
Quartzdiorite and granodorite		++	(+)+	+++ 40–55	(+)+	+		X	X	X	X	X				72,0	2,0	0,66
Biotite granite	St	++	++ 95	++ ~30		+	(+)+	X		X	X	X				71,9	1,9	0,89
Twomica and leucocratic granite	SB	++	++	++ 20			+			X	X			X	X	74,9	0,7	0,88
BRAJKOVAC																		
granodiorite	St	++	+	+++ 30–40	(+)	+		X	X	X	X	X	X			65,6	4,0	1,5
Aplitic granite	SB	++	++	++ ~15			+									71,7	2,7	0,24
CER																		
quartzmonzonite	W	++	++	+++ 36–40	+	+		X	X	X	X	X				65,4	4,2	1,18
Leucocratic granite	SB	++	++	++ ~10		(+)	+	X		X	X			X	X	72,0	1,3	0,78
Muscovite granite		+++	+++	++ 10–20			+				X			X		72,0	1,6	0,68
STRAŽANICA																		
Granodiorite	St	++	+	+++ 20–35		+	(+)	X	X	X	X	X				68,8	3,1	1,45
BOGATIĆ																		
Tonalite	?	++	(+)	+++	+	+		X	X	X	X	X				65,7	5,2	1,87
BORANJA																		
Granodiorite	St	++	++	+++ ~40	+	+		X	X	X	X	X				64,6	3,9	1,07

Contact metamorphism: St – strong; W – weak; SB – small bodies
* Kf with Or-content; *Pl with An-content

Signs: +++ main constituent; ++ abundant; + present; (rare).
X present accessory minerals

TABLE 2

K/Ar ages, Sr-isotope ratios and minimal residence age of crustal material (MRACM)

	K/Ar age, Ma	87Sr/86Sr	MRACM
BUKULJA, granodorite			
Biotite granite	Bi:27, wr:18+/-4	.7078	
Two mica and leucocratic granite	Bi, Mu, Kf:15–19	+/-2	
BRAJKOVAC			
granodiorite	wr:31–30	.7066	
	Bi:18+/-2	+/-4	
granodiorite porphyrite	wr:25+/-5		
CER			
quartz monzonite	Ho:30, wr:22	.708078	1 Ga
	Bi:15–18	0.708238	
leucocratic granite	Bi:14–18	.721239-	1.6 Ga
	Mu:15–18	.721313	
muscovite granite	Mu:15–18	.719714	
STRAŽANICA			
granodiorite	wr:18 +/-4	.706637	
	Bi:17.17 +/-.65	–	
	KF:18 +/-2	.707049	
	F:19.1 +/-0.8		
BOGATIĆ			
tonalite	F:36–37		
BORANJA		.70815	
granodiorite	wr, Bi, Ho:32.5–30	–	1 Ga
		.70864	

wr – whole rock
Bi – biotite
Ho – hornblende
Kf – K-feldpsar
Mu – muscovite
F – feldspar

tion of crustal melts which gave these rocks. By this intrusion the K-feldspar from the biotite granites ere probably rejuvenated and therefore the whole rock K/Ar data are close to the data of ages of the minerals in the muscovite bearing granites. The intrusions were of high thermal capacity: especially of the late S-granitic phase which induced strong contact metamorphosm, feldspatization of rocks at the contacts, and rejuvenated the older granites, but the intrusion which gave the biotite granites was probably too.

The granodiorite body of Brajkovac (*Fig. 1*) is built mainly of hornblende biotite granodiorite (with local transitions to tonalite). There are small intrusions of aplitic granites and very rare dykes of granodiotiteporphyrites, aplites and pegmatites. The

host rocks are strongly contact metamorphosed. The granodiotitic rocks are of I-type (Tables 1 and 2), and of magmatic (volcanic) arc affinity (*Fig. 3*). The obtained radiometric data by K/Ar method gave for the main variety a whole rock age of 30 Ma and for the granodioriteporphyrites 25 +/- 5 Ma but for the biotite only 18 +/- 2 Ma indicating probably a reheating to temperature above 300 °C.

The granitic massif of Cer (*Fig. 1*) is built mainly of quartz monzonite penetrated at margins and near the roof by leucocratic and muscovite granites. The quartz monzonite is a rock of I-type (Tables 1 and 2), magmatic (volcanic) arc affinity (*Fig. 3*), exhibiting however different isotopic age data: hornblende givers 30 Ma, the whole rocks analyses 22 Ma, but the biotite 15–18 Ma only. The muscovite granites and the leucocratic ones are of S-type (Tables 1 and 2), and of "collisional" affinity (*Fig. 3*). They have crystallized (according to the K/Ar ages of biotites and muscovites) 18 to 14 Ma age, i. e. simultaneously with the rejuvenation of biotite in the quartz monzonite. The contact metamorphism related to all these rocks is weak. The obtained data suggest that the quartz monzonite crystallized deeper in the crust 30 Ma ago but later mobilized probably because of the regional heating phenomenon and around 22 Ma ago reomorphic emplaced at the present place as a mass of low thermal capacity. Later on a second phase gave small masses of S-granites. These late melts because of small amounts could not induce additional metamorphism on surrounding rocks, but because of high fluid content (tourmaline, muscovite etc) locally produced a weak feldspatization of host rocks. The older quartz monzonites were by the intrusions of these younger granites heated up to temperatures only weakly exceeding 300 °C (biotite is rejuvenated, the feldspars mostly, but the hornblende is by all these processes influenced not at all).

It is very interesting that the minimal residence age of crustal rocks which contributed to the melts giving the quartz monzonites is about 1 Ga, but the material which gave the younger muscovite bearing granites 1.6 Ga.This may be explained either by a deepening of the magma generation area in the crust with time, what is not probable, or by intensive overthrusting of units existing in the deeper crust.

In the neighbourhood of the granitic massif of Cer occurs the small granodioritic body of Stražanica (*Fig. 1*). These granodiorites exhibit some peculiarities compared to most of the rocks of Cer. The are of I-type (Table 1 and 2), and of magmatic (volcanic) arc affinity (*Fig. 3*). The contact metamorphism associated to this granodiorite is strong. The granodiorites crystallized according to isotopic K/Ar data about 18 +/- 4 Ma ago (age of the whole rock and its mineral constituents). It is however probable that the magmas intruded earlier and that the obtained K/Ar age data 17–19 Ma indicate the time of a rejuvenation because of heating above 300 °C. Since in the Stražanica area does not occur any younger intrusions the rejuvenation, if happened, can be connected with a regional heating phenomenon only.

North of Cer, in the neighbourhood of Bogatić in some drillholes hornblende-biotite tonalities were found (*Fig. 1*). These tonalites belong to I-type granites (table 1); they crystallized about 37–36 Ma ago and are associated with Oligocene andesites and dacites.

DISCUSSION

The garnitic Bukulja–Cer–Motajica (BCM) belt differs from the granitic association of the Dinarides, Vardar zone and the Serbo-macedonian massif, i. e. from the central part of the Balkan peninsula (CBP) in the presence of the felsic varieties, geochemistry

and especially in the origin. The ones from the CBP are of the I-type and some originated about 30–32 Ma ago, others about 20 Ma ago but all in one phase (except the Zeljin massif) (KARAMATA *et al.* 1992). The granitic rocks of the BCM belt however originated in one two or even three phases, I-types as well as S-types are present and their development not uniform. Since some of the early phases in the BCM belt are similar to the CBP granitic rocks the data for the granodioritic rocks of Boranja, as their typical representative, are for comparison presented at the tables and at the *Fig. 3* (after KARAMATA 1955). The similarities of the first phases of Cer, Bukulja and of the Brajkovac granodiorite and Bogatić tonalite with the Boranja rocks are evident, the main intrusive phases at Bukulja and Stražanica are of I-type but differ in general from the Boranja anf other CBP granitic rocks, even from those of Miocene age, and the late phases of granite from Bukulja and Cer are of S-type, i. e. quite different from the CBP intrusives (even from the most similar Polumir granite).

Consequently the earlier phases of granitic magmatism of the BCM belt, which gave the hornblende bearing granodiorites of Brajkovac and Bukulja, the quartz diorite of Bukulja, the tonalite of Bogatić and probably the first deep crystallized equivalents of the quartz monzonites of Cer are similar to the granitic rocks of the CBP group, and can be considered as postcollisional I-granites of magmatic arc affinity identical to the granitic rocks widespread in the central part of the Balkan peninsula (Tables and *Fig. 3*). They were formed during Oligocene about 30 Ma ago as typical magmatic intrusions. The proof of their magmatic character is the strong contact metamorphism as well as the presence of associated volcanics (Brajkovac, Bogatić). At Bukulja the granodioritic rocks occur only as blocks in the younger granites and at Cer they were located deeper and later mobilized and reintruded, therefore their primary contact metamorphism can not be observed.

The biotite granitic rocks of I-type and magmatic arc affinity represent the main intrusions at Bukulja and Stražanica. At Stražanica there is a wide contact aureole, at Bukullja existed probably such one too, but is masked by the contact phenomena related to younger intrusions. The older K/Ar isotope age data (27–25 Ma) corresponding to Late Oligocene, were only determine at the biotite of the biotite granite of Bukulja. The whole rock age for the biotite granite of Bukullja, as well as the data for Stražanica biotite granodiorite (whole rock and feldspars) are around 18 Ma (Table 2). Here has to be mentioned that at Brajkovac granodioriteporphyrite dykes formed 25 Ma ago occur. These older aga data indicate a continuity of magmatic activity from about 30 Ma ago, when the early hornblende bearing granitic phase formed, until at least 25 Ma ago, when the biotite granite of Bukulja and the granodioriteporphyrite dykes of Brajkovac crystallized. Considering these data together with the almost identical Sr-isotope ratios (Table 2) of the early phases of the BCM magmatism (the Brajkovac granodiorite and the Cer quartz monzonite) and of the biotite granitic rocks (Bukulja and Stražanica) a consanguinity of both groups of rocks is indicated. The biotite bearing rocks, especially the one of Bukulja, can be the last most felsic differentiates of the deep seated magmas which gave the hornblende bearing granitic rocks. However the biotite granites could originate from magmas of the same provenance in deep crust an Upper Mantle as those which gave the hornblende bearing rocks too, or also, most probably the Stražanica granodiorite, from magma deriving by strong remelting of older granitic rocks connected with the heating phenomenon along the BCM belt. To this event also can be connected the mobilization and partial melting of the quartz monzonitic masses below Cer and their emplacement at the present position. The age of this mobilization and melting phenomenon can not be exactly determined. It was after the first phase i. e. after 25 Ma ago, but before about 18 Ma ago when the muscovite bearing granites intruded the quartz

monzonites of Cer and rejuvenated the minerals of the older rock of Cer and Stražanica. This event is probably simultaneous with the Miocene phase of the granitic magmatism in the CBP, which happened about 20 Ma ago.

After a short interruption a new phase of magmatic activity originating however from a quite different crystal level gave the S-type muscovite bearing rocks of Bukulja, Cer and small masses at Brajkovac (?) (muscovite-biotite, muscovite, muscovite +/- tourmalline +/- garnet granites). These granites were intruded about 18 Ma ago and crystallized until about 15 Ma ago. They intruded the older granitic rocks added some contact alterations at the surrounding rocks, especially feldspatization and tourmalinization, and together with a related low intensity heat flow along the BCM belt, rejuvenated the minerals of the older granites (all minerals of the Bukulja biotite granite, biotite of the Brajkovac granodiorite and the Cer quartz monzonite as well as of the Stražanica granodiorite). The Sr-isotope ratios of these rocks proove that they have crystallized from magmas originated by melting of metasedimentary rocks and probably some very old granitic rocks in the deep crust.

CONCLUSION

The final conclusions about the development of the granitic magmatism in the Bukulja–Cer–Motajica belt can be summarized as follows:

1) About 30 Ma ago I-type magmas were intruded and their differentiation and crystallization lasted until about 25 Ma ago. They gave the hornblende bearing granodiorites of Brajkovac and Bukulja, the deep seated facies of the quartz monzonites of Cer and the tonalites of Bogatić as the first and biotite granites of Bukulja and dyke rocks of Brajkovac as final products.

2) Around 20 Ma ago a strong heating happened along the BCM belt, inducing mobilization and partial to more advanced elting of the previously formed granitic rocks. The mobilized and only initially melted rocks rose diapirically and gave reomorphic intrusions of quartz monzonites at Cer. The melts originated by more advanced melting wrere separated from the restite and intruded as magmas which gave the biotite granodiorite of Stražanica. These rocks are of I-type.

3) About 18 Ma ago were intruded S-type magmas originated by melting of metasedimentary and old felsic units in the crust because of strong heat flow in the roots of the BCM belt. These magmas were intruded along the BCM belt, mostly at the earlier preferred sites where the older granites were intruded earlier and gave muscovite bearing granites of Bukulja. Cer and probably Brajkovac.

4) The magmatism of the first and the second phase is widespread all around the central part of the Balkan peninsula and all over similar rocks are present. It is of postcollisional character referred to compressional processes in the crust and probably disturbances in the uppermost Mantle. The youngest S-type magmatism is restricted (with exception of the similar Polumir granite) to the BCM belt. It is confined to southernmost fracture zone of the set of faults at the southern margin of the Pannonian basin. Along this deep fault the heat advanced upwards and melted the metasedimentary (+/- very old granitic) rocks. It has to be stressed that there was not any collision even not any important vertical or horizontal displacement at that time in this region and therefore the "collisional" affinity of these granites is not related to any collision bu to the character of metamorphic rocks which gave the melts. Simillar geological conditions did not exist southwards in the CBP and for this reason such rocks only in the BCM belt occur.

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AGES AND INTENSITIES OF METAMORPHIC PROCESSES IN THE BATOČINA AREA, SERBO–MACEDONIAN MASSIF

KAD. BALOGH, É. SVINGOR*

Institute of Nuclear Research of Hungarian Academy of Science

V. CVETKOVIĆ**

Faculty of Mining and Geology, University of Belgrade

ABSTRACT

The rocks of the polymetamorphic complex of Batočina were studied by the Rb/Sr and K/Ar isotopic methods in order to determine the age and intensity of metamorphic phases.

Rb/Sr data of eight whole rock samples are arranged along two regression lines defining model ages of 709±51 Ma and 488±19 Ma. The older age is chronologically meaningless since RL–I is a mixing line. The younger datum of 488±19 Ma gives th age of the oldest recognized metamorphism.

Several K/Ar ages on amphiboles indicate Hercynian age, while on the basis of K/Ar ages on muscovite and K-feldspar 0.30±0.13 mm/year uplift rate was obtained for the time interval from 150.6 Ma to 127.3 Ma.

INTRODUCTION

The metamorphic series of the Batočina area, representing the northernmost exposed parts of Serbo-Macedonian Massif, consists of various metamorphic rocks originating from different protolites. The petrological and geological significance of these rocks enforced this investigation to explain some problems concerning the age of different phases of metamorphism. The only existing radiometric data for this terrain (BALOGH and CVETKOVIĆ, 1992 in print) closely agree with the earlier investigations in the central parts of the Serbo-Macedonian Massif by DIMITRIJEVIĆ (1971) and MILOVANOVIĆ (1990). According to the above mentioned data as well as to the geological evidences (PETROVIĆ and KARAMATA, 1970) the latest metamorphic phases were Harcynian (275 Ma – hornblende analysis) and Alpine (127–160 Ma – mica and potassium feldspar analysis).

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* H–4001 Debrecen, P. O. Box. 51.
** YU–11000 Belgrade, Djušina 7.

GEOLOGICAL SETTING

The Batočina series (about 900 m thick) represents the furthermost northern parts of the "Lower complex" of the Serbo–Macedonian Massif (DIMITRIJEVIĆ, 1959) – a terrain which plays an important role in the geological structure of Yugoslavia and the central part of the Balkan peninsula (*Fig. 1*). It consists of various types of gneisses, micaschists, amphibolites, marbles, quartzites and other schists. Earlier investigations (BOUE, 1836; HERDER, 1846; MILOJKOVIĆ, 1892; UROŠEVIĆ, 1912; DIMITRIJEVIĆ, 1950; CMILJANIĆ and DJORDJEVIĆ, 1969; KALENIĆ *et al.*, 1975; VUJISIĆ *et al.*, 1977) suggested a prograde metamorphism corresponding to the epidot-amphibolite to amphibolite facies. According to petrological and especially geochemical evidences the complex in a whole represent a metamorphosed deep-water sedimentary unit of Riphean-Cambrian age (VUJISIĆ *et al.*, 1977) with late shallow-water facies, while amphibolites correspond to the ancient subalkaline within-plate tholeiites (CVETKOVIĆ, 1992). On the basis of chemical characteristics of the minerals in the equilibria CVETKOVIĆ (1992) has concluded that metamorphism occurred at a temperature between 550 and 600°C and a pressure at about 6 kb. The mentioned values are strongly supported by the mineral assemblage (garnet – staurolite – kyanite in gneisses and micaschists and epidote – hornblende – garnet in amphibolites.)

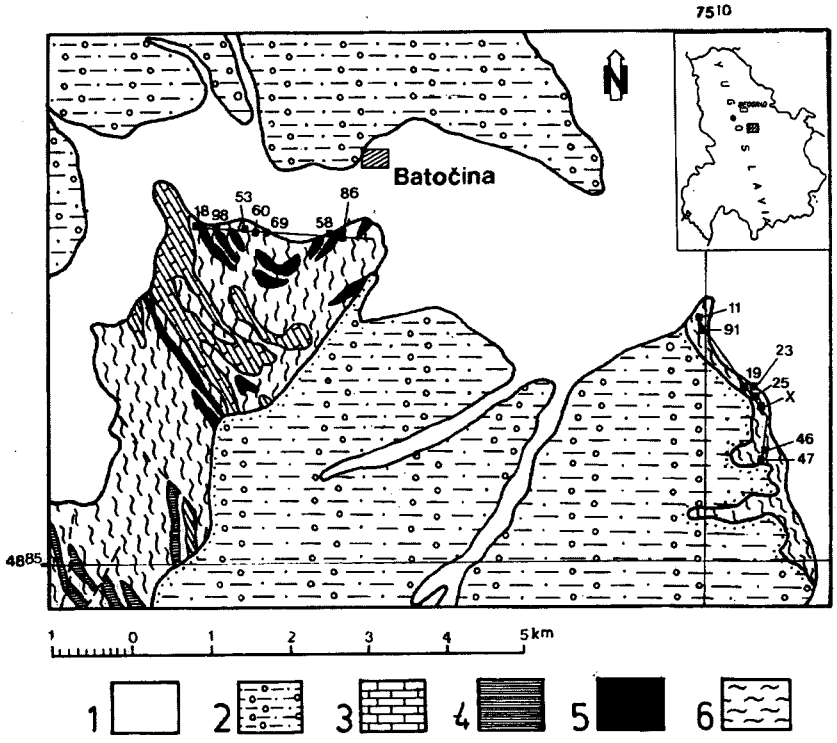

Fig. 1. Geological sketch of the Batočina region
Legend: 1 – Quaternary;
2 – Lower and Upper Miocene sediments;
3– Marbles and dolomite marbles;
4 – Quartzites;
5 – Amphibolites;
6 – Gneisses and misaschists.

PETROGRAPHY AND SAMPLING

Finegrained biotite and coarsegrained two-mica paragneisses, which with micaschists together usually form layer units up to 15 m thickness at the best exposures, and augen-ortho(?) gneisses were distinguished. Paragneisses and micaschists often show transition while augen-gneissen are usually interlayered with host metasediments or form, rather seldom, small discordant bodies always with sharp contacts. The amphibolites are relatively scarce in comparison to the metapsammitic and metapelitic country rocks; they occur as lens shaped bodies originated by boudinage from 0.5 m to 15 m in length, usually 0.5–1m thick, parallel to the foliation.

Both finegrained biotite and coarsegrained two-mica gneisses are of granoblastic to lepidoblastic texture and of distinctive schistosity. Elements of finegrained blastopsammitic texture also occur. They consist of quartz, plagiclase (32–52% An), rare alkali feldspar, fragments of quartzites, gneisses, granodiorite and rarely phylites, biotite, muscovite, garnet, apatite, tourmaline and opaque minerals. Secondary minerals are chlorite, sericite and seldom epidote. In some varieties close to micaschists rare euhedral staurolite and kyanite crystals were also observed.

Otherwise, the augen-gneisses show lepidoblastic and granoblastic texture and augens, sometimes even amygdaloidal strusture. They are composed of quartz, alkali feldspar (microcline, ortoclase and albite often with myrmekite), muscovite, biotite, apatite, zircon, opaque minerals and secondary chlorite and sericite.

The Batočina amphibolites can be divided into two groups well distinguishable in the field:

– garnet-amphibolites mainly of massive structure and nematoblastic to granoblastic texture essentially consisting of hornblende, plagioclase, garnet, sphene, zoisite, epidote, ilmenite and quartz,

– garnet -biotite amphibolites prevailing of schistose structure with biotite as an abundant constituent, compared with the above mentioned amphibolites, very rare zoisite and sphene.

Eight samples of paragneisses and micaschists, two of augen-gneisses and four samples of amphibolites have been collected across the two exposed profile (*Fig. 1*). Sample distances range from several tens of meters to a few decimeters depending on rock unit homogenity. The main petrographic features as well as mineral compositions of the analyzed samples are presented in Table 1.

EXPERIMENTAL METHODS

Rb/Sr Method

Rb ans Sr were measured by stable isotope dilution methods. The spike solution was added to the weighed sample at the beginning of the procedure. Following this the samples were taken into solution in a mixture of hydrofluoric and perchloric acids and subsequently converted to chlorides. Sr was separated and purified on a cation exchange column using Dowex 50W resin. All Rb and Sr analyses were made with a modified MI-1309 mass spectrometer. At least 50 scans of the Sr mass spectrum were used in the final data compilation. During the period of this study the average $^{87}Sr/^{86}Sr$ obtained for EIMER and AMAND Sr-standard ($^{87}Sr/^{86}Sr=0.70802\pm2$) was $0.7079\pm2(2\sigma$ of the mean, 10 analyses). To control also the chemical procedure, $^{87}Sr/^{86}Sr$ ratio of a basalt sample from Diszel (W-Hungary) was measured, the ob-

TABLE 1

Main petrographic features and mineral composition of the analyzed metamorphic rocks from the Batočina complex

No	Rock Type	Texture	Mineral composition													
			Q	Af	Pl	Mu	Bi	Ho	Gt	St	Ky	Ap	Sph	Ep	OM	To
BČ-X	TMG	lpb, gbl, pbl	+	*	+	+	+	–	+	–	–	*	–	–	*	-
BČ-10	FBG	lpb, gbl	+	–	+	–	+	–	+	*	–	*	–	–	*	-
BČ-11	AOG	gbl, lpb	+	+	–	*	+	–	–	–	–	*	–	–	*	-
BČ-18	GA	nmb, gbl	–	–	*	–	–	+	+	–	–	*	+	*	*	-
BČ-19	FBG	bpsm	+	–	+	–	+	–	+	–	–	*	–	–	*	-
BČ-23	TMG	lpb, gbl, pbl	+	–	+	+	+	–	*	*	–	*	–	–	*	-
BČ-25	TMG	lpb, gbl, pbl	+	*	+	+	+	–	+	–	–	*	–	–	*	-
BČ-46	FBG	lpb, gbl	+	*	+	–	+	–	+	*	–	*	–	–	–	-
BČ-47	TMG	lpb, gbl, pbl	+	–	+	+	+	–	+	*	–	*	–	–	*	-
BČ-53	AOG	gbl, lpb	+	+	–	*	+	–	–	–	–	*	–	–	*	-
BČ-58	GBA	nmb, gbl, lpb	*	–	*	–	+	+	+	–	–	*	–	+	*	-
BČ-60	MSC, TMG	lpb, gbl	+	–	*	+	+	–	+	*	*	*	–	–	–	-
BČ-69	MSC	lpb, gbl, pbl	+	–	–	+	+	–	*	–	–	–	–	–	–	*
BČ-86	GBA	nmb, gbl, lpb	*	–	+	–	*	+	+	–	–	*	–	*	*	-
BČ-91	MSC	lpb, gbl, pbl	+	–	*	+	+	–	+	*	*	*	–	–	*	-
BČ-98	GA	nmb, gbl, pbl	*	–	+	–	–	+	+	–	–	*	+	*	*	-

EXPLANATIONS:
ROCK TYPE: FGB - Finegrained biotite gneisses; TMG – Two-mica gneisses; AOG – Augen orto gneisses; MSC - micaschists; GA – garnet amphibolites; GBA – Barnet biotite amphibolites;
TEXTURES: lpb – lepidoblastic; gbl – granoblastic; nmb nematoblastic: pbl – porphyroblastic; bpsm – blastopsammitic;
MINERAL COMPOSITON: Q – quartz; Af – alkali feldspar; Pl – plagioclase: Mu muscovite: Bi – biotite; Ho – hornblende; Gt – garnet; St – staurolite; Ky – kyanite; Ap – apatite; Sph – sphene; Ep – epidote (zoisite); OM – opaque minerals; Tou – tourmaline;
SIGNES: + abundant; * rare; - absent.

tained ratio was 0.7042±6. This sample was analysed in Royal Holloway and Bedford New College (RHBNC), the measured $^{87}Sr/^{86}Sr$ ratio was 0.70409 (EMBEY-ISZETIN *et al.*, 1993).

From cumulative data we estimate that two standard deviations of the mean for $^{87}Sr/^{86}Sr$ is 2.0%. This value was used for purposes of regression analysis. The individual $^{87}Sr/^{86}Sr$ run precision was used for each data point because this uncertainty varied considerably for differrent analyses. The errors given for age and intercept in Table 2 were estimated from these values, the errors given in brackets were calculated from the scatter of data points on the lines.

K/Ar method

Measurement of K/Ar ages was performed in the Institute of Nuclear Research of the Hungarian Academy of Sciences (ATOMKI), Debrecen, Hungary. The samples were first crushed to a size required by the mineral separation (0,063–0,315 mm, magnetic separation, heavy liquids and shaking off on a sheet were used for mineral separation) then a part of the sample was selected and pulverized for K determination.

An argon extraction line and a mass spectrometer, both designes and built in ATOMKI, were used for the Ar measurement.The rock was degassed by high frequency induction heating, the usual getter materials (titanium sponge, CuO, SAES getter and cold traps) were used for cleaning Ar. the ^{38}Ar spike was introduced to the system from a gas-pipette before the degassing was started. The cleaned Ar was directly introduced into the mass spectrometer. The mass spectrometer was a magnetic sector type of 150 mm radius and 90° deflection, it was operated in static mode. Recording and evaluation of Ar spectrum was controlled by a microcomputer.

0.1 g of the pulverized samples were digested in HF with the addition of some sulphuric and perchloric acids. The digested sample was dissolved in 100 ml 0.25 mol/l HCl and after a fivefold dilution 100 ppm Na and 100 ppm Li were added as buffer and internal standard. K concentration was measured with a digitized flame photometer of OE-85 type manufactured in Hungary.

The interlaboratory standards Asia 1/65 and GL-O and atmospheric Ar were used for controlling the measurements. Errors were calculated with estimating 1% error (1σ) for the isotope ratios and 3% for the potassium concentrations. Ages were calculated with the constants suggested by STEIGER and JÄGER (1977). Details of the instruments, the applied methods and results of calibration have been described elsewhere (BALOGH, 1985; ODIN, 1982).

DISCUSSION

Rb/Sr ages

Eight whole rock samples were mneasured by Rb/Sr method. Six of them were paragneisses, one was an amphibolite (BČ-98) and one was a tourmaline bearing micaschist (BČ–69). The isotopic ratios, Rb and Sr concentration data, model ages, as well as initial ratios and ages calculated from regression lines are presented in Table 2. The samples have basic characters, except the sample BČ–69, so we have calculated the model ages with assumed initial ratio of $(^{87}Sr/^{86}Sr)_i=0.703$, characteristic for basaltic magmas,.

As it can be seen in Table 2, the model ages form two groups: 455–511 Ma and 699–841 Ma. Accordingly, the eight samples do not define a single regression line, but

TABLE 2

Rb/Sr ages on whole rocks samples from the Batočina area

Sample		$^{87}Sr/^{86}Sr$	Sr ppm	Rb ppm	$^{87}Sr/^{86}Sr$ atomic	Model Age (Ma)	Isochron Age (Ma)	$(^{87}Sr/^{86}Sr)_i$
BČ-19	paragneiss	0.7072± 7	264.63	60.29	0.6361	467± 85		
BČ-23	paragneiss	0.7059±12	292.48	46.63	0.4451	455±206	488±19(10)	0.7030±6
BČ-46	peregneiss	0.7094± 8	265.31	83.99	0.8840	511± 73	RL-II	
BČ-69	micaschist	0.7406± 9	51.94	105.94	5.7139	486± 21		
BČ-10	paragneiss	0.7188±11	172.85	92.94	1.5028	738± 67		
BČ-47	paragneiss	0.7149± 6	226.97	87.98	1.0830	769± 56	709±51(41)	0.7036±6
BČ-60	paragneiss	0.7162±11	200.50	94.60	1.3184	699± 71	RL-I	
BČ-98	amphibolite	0.7068± 4	313.01	35.50	0.3166	841±112		

two definitely different ones (*Fig. 2*). The data points lie on the individual lines within the analytical uncertainties. The age calculated from regression line fitted on samples BČ-10, BČ-47, BČ-60 and BČ-98 (RL-I) is 709±51(41) Ma with an initial ratio of 0.7036±6, the age defined by the other regression line (RL-II, BČ-19, BČ-23, BČ-46, BČ-69) is 488±19(10) Ma with an intercept of 0.7030 ±6.

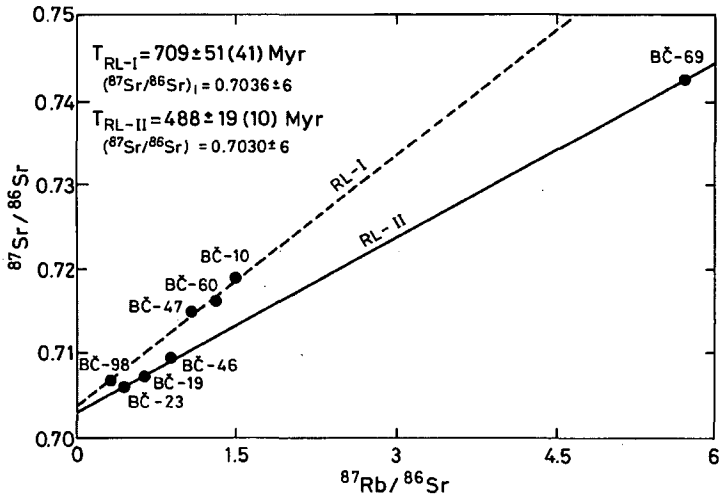

Fig. 2. Rb-Sr regression lines for whole rocks samples from Batočina region.

A regresion line may be an isochron dating a geological event or a meaningless mixing line. In this case we have two regression lines, RL-I and RL-II, so the following different possibilities will be considered:

– RL-I and RL-II are isochrons.

RL-I may give tha age of sedimentation. This is unlikely, because of the point of BČ-98, which belongs to an amphibolite, and because of the low initial ratio, which is characteristic for basic materials and not for sedimentary rocks. It is more probable, that it gives the time of metamorphism having formed these gneisses. Fluids coming from the upper mantle or the lower crust with high Sr content and low Sr isotope ratio might have resulted in this isotope homogenization.

If RL-I is an isochron, and the rocks forming the two regression lines are not completely independent, RL-II can give neither the age of sedimentation, not the time of their first metamorphism. It can date a second event, which was intensive enough to form homogeneous isotope composition but did not have any influence to the rocks forming the other isochron. For this we have to suppose again the perfect independence of the two groups. The intercepts of the two lines agree within the analytical uncertainties, and this argues for their relationship. So it is unlikely, that both of the two regression lines are isochrons.

– RL-I is an isochron, RL-II is a mixing line.

In this case the isotope ratios and Rb and/or Sr concentrations of the rocks lying on RL-II regression line were modified by a later event, which was not strong enough to

form a new isotope equilibrium in a large scale, the regression age has no meaning. Sample BČ-69 is an argument against it. It has higher $^{87}Sr/^{86}Sr$ ratio, lower Sr and a bit higher Rb content than the others, so its model age is relatively insensitive to the assumed initial ratio giving the most probable time of the metamorphism. Since this sample lies on RL-II, there is an alternative possibility for explaining the results:

– RL-I is a mixing line, RL-II is an isochron.

In this case the regression line RL-II has real menaning giving the time of metamorphism having formed these gneisses. The effect of fluids of low Sr isotope ratio (characteristic for fluids from upper mantle or lower crust) on the old sedimentary rocks might result in complete isotope homogenization in a part of a large area, but not necessarily over the whole area. Where the isotope homogenization wasn't complete, a regression line was produced with a virtual age, so the mixing line will show an older age then the time of metamorphism. If RL-II dates the time of metamorphism, this event might form the RL-I mixing line for a part of the rocks of the area. It can explain the similarity of the two intercepts, and even the fact, that RL-I has a bit higher initial ratio (within the analytical uncertainties). The better fitting of the points on line RL-II also supports this assumption. Sample BČ-69 belonging to RL-II is an argument for it, too.

– RL-I and RL-II are both mixing lines.

We have to consider this possibility as well. In this case the regression lines were formed by later events and we can't say anything about the time of metamorphism. But different events can't form two mixing lines with the same intercepts on the same area, they would cause only the scatter of the points.

On the basis of theoretical interpretation of Rb/Sr data we may say: Independently of the meaning of the two regression lines (isochron or mixing line) the values of the intercepts show that the sedimentary rocks had a basic character, but even supposing on extremely basic character for them the fluids effecting on the sediments during the metamorphism would have had $^{87}Sr/^{86}Sr$ ratio not higher than 0.703, therefore they must have originated from the upper mantle or the lower crust. The similarity of their intercepts suggests that the sediments over the studied area had similar Sr content and isotope composition and/or the fluids effecting them were the same. If the fluids were the same, we can exclude the possibilities of having two isochrons or two mixing lines. Among the above mentioned interpretations of the regression lines the third one is the most convincing: RL-II regression line has a real geological meaning, giving 488±19(10) Ma for the time of metamorphism and RL-I is a meaningless mixing line formed by the same event. At the same time this means, that the $^{87}Sr/^{86}Sr$ ratio of sediments was a little higher than the initial ratios defined by RL-I and RL-II. This moment we have no data contradicting this interpretation.

K/Ar ages

K/Ar ages are summerized in Table 3. Ages range from 965 Ma to 127.3 Ma, clearly, they were overprinted by secondary thermal events. In metamorphic rocks K/Ar ages are usually connected with the time, when the rock cooled below the closure temperature of the dated minerals. The closure age of minerals shows either the time when the area was lifted up, or the time of a reheating event; both can be much younger than the age of metamorphism. On the other hand, in metamorphic rocks at elevated temperatures radiogenic Ar may diffuse out of the minerals and a radiogenic Ar atmosphere may evolve at the mineral boundaries. The radiogenic Ar may diffuse in the minerals, and the K/Ar age of low K minerals may increase considerably this way.

TABLE 3

K/Ar ages on minerals from the Batočina area

Sample (K/Ar N°)		Dated mineral	K %	^{40}Ar(rad) 10^{-5}ccSTP/g	^{40}Ar(rad) ^{40}Ar(tot)	36Ar(atm) 10^{-9}ccSTP/g	Age Ma
Paragneisses							
BČ-X	(2451)	muscovite	6.28	3.705	0.82		145.9 ± 5.7
		plagioclase	0.21	0.375	0.22		417 ± 46
				0.360	0.65		402 ± 44
BČ-10	(2768)	muscovite	3.56	2.121	0.89		147.2 ± 5.9
BČ-19	(3039)	white mica	1.63	1.135	0.43		171 ± 7.7
BČ-25	(2450)	mica	4.88	4.802	0.83		237 ± 9.0
BČ-47	(2770)	biotite	5.74	4.034	0.91		172.4 ± 6.9
		muscovite	6.80	4.030	0.93		146.4 ± 5.8
		plagioclase	0.16	0.170	0.61		259 ± 27
BČ-69	(3043)	muscovite	8.25	5.177	0.90		154.7 ± 5.8
BČ-91	(2402)	biotite	6.22	3.649	0.84		145.1 ± 5.5
		muscovite	6.23	4.021	0.89		159.0 ± 6.0
Amphibolites							
BČ-18	(2452)	amphibole	0.19	0.913	0.82	6.8	938 ± 103
				0.946	0.85	5.7	965 ± 106
BČ-58	(2771)	amphibole	0.39	1.306	0.63	26.0	711 ± 43
BČ-86	(2401)	amphibole	0.52	0.595	0.78	5.7	276 ± 11
BČ-98	(2769)	amphibole	0.62	0.795	0.72	10.5	303 ± 17
	(2449)	amphibole	0.58	0.834	0.61	18.0	342 ± 19
	(2769)	biotite	4.63	3.061	0.52		162.7 ± 6.6
	(2769)	plagioclase	0.17	0.500	0.65		639 ± 66
Orthogneisses							
BČ-11	(2403)	K-feldspar	8.43	4.283	0.95		127.3 ± 4.8
BČ-53	(2767)	K-feldspar	5.89	3.282	0.86		138.0 ± 5.5
	(2453)	feldspar	3.01	2.411	0.85		195.6 ± 7.8

In the Batočina region oldest ages were measured on amphiboles (the mineral with highest closure temperature), and the youngest ages on K-feldspar, the mineral that loses Ar at the lowest temperature. The closure temperature of plagioclase is variable, it depends on the crystalline structure of the dated mineral, but it is always lower than that of muscovite. Several old ages in Table 3 (BČ-X, BČ-47 and BČ-98) prove, that radiogenic Ar was incorporated by plagioclase during metamorphism, therefore plagioclase K/Ar ages are older than the time when cooled below their closure temperature.

The concentration of muscovite, biotite and K-feldspar ages in the 172.4–127.3 Ma interval indicates a thermal event (uplift or reheating of the area) during this period. PURDY and JÄGER (1976) determined 350±50°C for the closure temperature of muscovite. HARRISON and MCDOUGALL (1982) and HARRISON *et al.* (1979) calculated 130°C for microcline and 160°C for K-feldspar. The closure temperature of biotite is less certain, 300°C, 300±50°C, 300–345°C and 280±40°C were given for it by BERGER and YORK (1979), PURDY and JÄGER (1976), DALLMEYER (1978) and HARRISON and MCDOUGALL (1980a). However, the analysis by CLIFF (1985) suggests, that the closure temperature of biotite may be a function of the geological enviromnent and can be higher than that of the muscovite. RODDICK *et al.* (1980) have shown that biotite incorporates excess Ar during metamorphic cooling and this may result anomalously old ages. Due to this uncertainties we shall rely on the muscovites and K-feldspar ages.

Five muscovite sample from the paragneiss (BČ-X, BČ-10, BČ-47, BČ-69 and BČ-91) resulted in an average age of 150.6±5.8 Ma. The error of average age is similar to that of the individual ages, therefore at the present state of our study we disregard the scatter of muscovite ages and accept the average for the time of cooling below 350± 50 °C.

Two K-feldspar sample were dated. Excess Ar is likely present in BČ-53, since the feldspar with lower K content from this rock resulted in a remarkably older age. Therefore the age of BČ-11 (127.3±4.8 Ma) is accepted as the time when the rock cooled below 145± °C, i. e. the average closure temperature for K-feldspar. The muscovite and K-feldpsar ages and closure temperatures result in 8.8±3.6 °C/Ma cooling rate and assuming 30±5 °C geothermal gradient 0.30±0.13 mm/year uplift rate is obtained for time interval from 150.6 Ma to 127.3 Ma.

K/Ar ages on aphiboles from the amphibolites ranges from 965 to 276 Ma. They show that the Upper Jurassic – Lower Cretaceous thermal event didn't reset the K-Ar system of amphiboles. The closure temperature of amphibole is about 500°C (480°C by DALLMEYER (1978); 400–540°C, depending on the grain size and cooling rate, by HARRISON and MCDOUGALL (1980b), this sets an upper limit for the temperature of the Mesozoic thermal event.

The old ages obtained on amphiboles from BČ-18 and BČ-58 may be due to the incorporation of excess Ar or they may reflect the time of older metamorphic events. Since K/Ar age of amphibole cannot be older than the Rb/Sr whole rock age, we tried to demonstrate the presence of excess Ar in BČ-18 according to the method suggested by FECHTIG and KALBITZER (1966). If excess Ar diffused in the amphibole during a metamorphic event, and didn't reach equilibrium, then this Ar will be more easily released during a stepwise degassing experiment. This will result in an Arrhenius diagram as a bias from the straight line, the slope and the activation energy will be less for the lower temperature steps. *Fig. 3* shows that this is not so, the points are arranged along a straight line and Ar release is characterized by a single activations energy. This may indicate either that excess Ar is missing from the sample, or that incorporated excess Ar equilibrated in the mineral and it cannot be distinguished from radiogenic

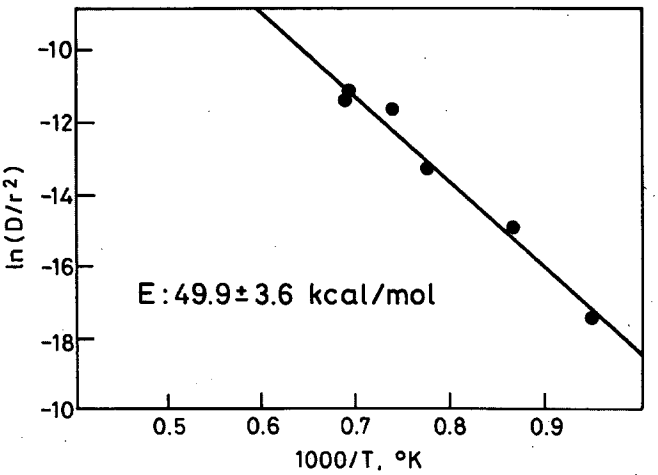

Fig. 3. Arrhenius diagram of Ar release from Amphibole of BČ-18.

Ar. The first possibility is discarded in the light of Rb/Sr ages, the second needs further discussion.

LEE (1993) demonstrated, that Ar release from hornblende in vacuo is not a volume diffusion process, but it is controlled by the internal stress of the mineral, caused by the oxidation of Fe^{2+} parallel with dehydrogenation. This mechanism of Ar release is termed short circuit diffusion; LEE obtained 45 kcal/mol activation energy for it, it is simillar to the value measured by us. Stimulation of Ar release by Fe^{2+} oxidation explains why Mg-rich amphibole loses Ar in vacuo at higher temperatures than Fe-rich samples do. Ar release from BČ-18 by short circuit diffusion is supported by the lower diffusion parameter at 1461 °C than at 1452 °C (Table 4). However, since radiogenic Ar was observed first at 738 °C, the presence of weakly bound Ar in this sample is very unlikely, i. e. radiogenic ad excess Ar cannot be distinguished. This involves, that definite chronologic meaning cannot be assigned to the old ages of samples BČ-18 and BČ-58.

TABLE 4

Stepwise release of $^{40}Ar(rad)$ from amphibole from BČ-18

Tempereture of heating		Released $^{40}Ar(rad)$	Diffusion parameter	Released fraction of $^{40}Ar(rad)$
°C	°K	ccSTP/g	D/r^2	
680	953			
784	1057	$2.395*10^{-7}$	$2.455*10^{-8}$	0.0258
885	1158	$6.188*10^{-6}$	$3.020*10^{-7}$	0.0666
996	1269	$1.175*10^{-6}$	$1.648*10^{-6}$	0.1264
1087	1360	$2.261*10^{-6}$	$8.538-10^{-6}$	0.2432
1179	1452	$1.778*10^{-6}$	$1.493*10^{-5}$	0.1913
1188	1461	$7.792*10^{-7}$	$1.075*10^{-5}$	0.0838

Amphiboles from BČ-86 and BČ-98 resulted in Hercynian ages. Since the presence of excess Ar, in view of the old plagioclase age, is evident in BČ-98, the obtained amphibole ages may be older than their closure ages. As FOLAND (1979) demonstrated, the concentration and isotopic composition of Ar atmosphere can be rather heterogeneous in a metamorphic terrain within small distances. The potassium content of the dated amphiboles is similar, therefore the application of isochron methods doesn't help to solve the question. If excess Ar was incorporated with uniform isotropic composition (in our case it is not so) than the amount of excess Ar will be proportional to the atmospheric Ar content. In the amphibole samples the Ar(atm) content. vary much more than the ages, and this indicates that in the samples the amount of radiogenic Ar is much more than that of the excess Ar. This is the only argument for a Hercynian event in the studied area.

CONCLUSIONS

The metamorphic complex of Batočina (composed of paragneisses, orthogneisses, micaschists and amphibolites was submitted to metamorphic and thermal events several times.

The Rb/Sr and K/Ar analysis enabled to distinguish the following processes:

1) The sedimentary rocks (unmetemorphosed before?) were metamorphosed above 500°C 488± 19 Ma ago during the Caledonian orogeny. Fluids from the lower crust or upper mantle equilibrated the Sr isotopic ratios.

2) Hercynian orogeny is indicated by the K/Ar ages of several amphiboles.

3) A thermal event resetting muscovite, biotite and K-feldspar ages (> 350±50°C), but not affecting amphiboles occurred before 150.6±5.8 Ma (Upper Jurassic), and the area cooled below 145±15 °C 127.34.8 Ma age (Lower Cretaceous). A cooling rate of 8.8±3.6°C/my was calculated from the muscovite and K-feldspar ages for the time interval from 150.6 Ma to 127.3 Ma.

Comparing our results with existing U/Pb ages of 270 Ma on zircon by DIMITRIJEVIĆ (1971) from the Bujanovac granite and K/Ar age of 140 Ma on hornblende by MILOVANOVIĆ (1990) from between Tulare and Lebane it can be concluded that in the southern region of the SMM either the intensity of both the Variscan and Upper Jurassic metamorphism was higher than in the Batočina area, or deeper levels of the SMM were exposed in the south.

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PETROGRAPHICAL CHARACTERISTICS OF DITRÓ (OROTVA) DIORITES, EASTERN CARPATHIANS, TRANSYLVANIA (ROMANIA)

E. PÁL MOLNÁR*

Department of Mineralogy, Geochemistry and Petrography, Attila József University

ABSTRACT

One of the most important rock groups outcropping in the NW part of the Ditró syenite massif, north of the Orotva creek and between the Csibi Jakab and Tászok creeks is the group of the dioritic rocks. On the basis of the mineralogical composition, color index (M) and the textural characteristics, this rock group can be divided into the following rock types: (1) meladiorites:meladiorites with textural ordering, meladiorites without textural ordering, schistose meladiorites; (2) normal diorites: diorites with textural ordering, diorites without textural ordering, feldspar schlieren diorites,diorites with feldspar aggregates; (3) leucodiorites: leucodiorites with textural ordering, leucodiorites without textural ordering.

As the 1:5000 geological map of the area shows, these diorites do not form layer like and lenticular bodies (as some researchers have suggested), but they are irregular rock bodies showing gradual transition and joggle toward the neighboring rocks. Diorites represent a connection between the hornblendites and the metamorphic rocks as well as between the syenites, monzonites and granites. To form this tectonic setting two (spatially equivalent but temporally different) magmatic mass had to be intruded.

INTRODUCTION

The syenite massif of Ditró (46°48′ N, 25° 30′ E) is situated in the S-SW part of the Gyergyó Alps belonging to the Eastern Carpathians. Diameter of its surface are 19 km and 14 km in NW and SE directions, respectively; its area is 225 km^2 including the bordering zones as well (*Fig. 1*).

In the NW part of the syenite massif, north of the Orotva creek and between the Csibi Jakab and Tászok creeks (*Fig. 1*) natural and artificial exposures, mines and drillings made an accurate geological mapping possible. The approximately 6 km^2 area between a major fault along the Orotva creek and the massif as well as the contact of the neighboring metamorphic rocks was mapped. Base of the 1:5000 geological map (*Fig. 2*) was a geological map made by ZÓLYA in 1986, which partly touches the area. During the field work, beside the accurate representation of the petrographical and tectonic conditions of the area, ultrabasic and intermediate rocks were studied in particular because I believe that genetics and petrogenetical classification of these rocks can bring genetics of the whole massif to light (PÁL MOLNÁR 1992).

In the mapped area, CODARCEA *et al.* (1957) divided the rocks outcropping in the Orotva valley from the west to the east into four rock complexes: (1) Complex of the Diorite-hornblendite Rocks; (2) Complex of the Syenitoid Rocks; (3) Complex of the Granitoid Rocks; and (4) Complex of the Dike-rocks. In their opinion, rock on the

* H–6701 Szeged, P.O. Box 651, Hungary

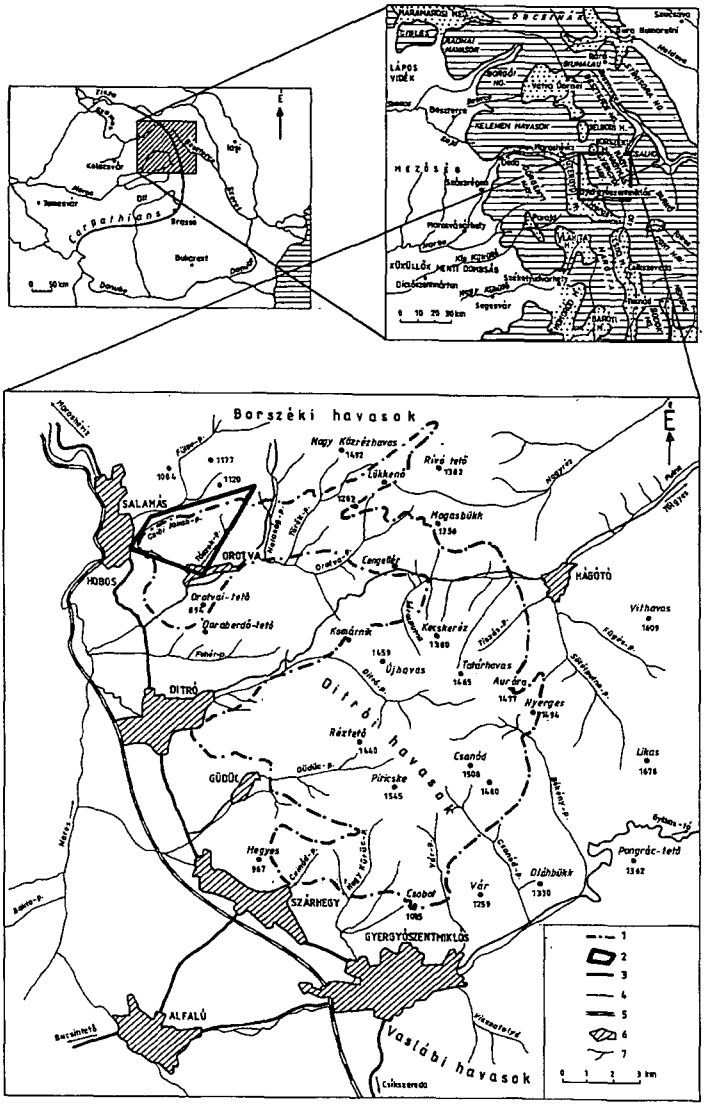

Fig. 1. Geographic environment of the Gyergyó Alps (Transylvania, Romania)
1. Boundary line of the Ditró Syenite Massif; 2. Study area; 3. Modernized road; 4. Non-modernized road; 5. Railway; 6. Settlement (town); 7. River (creek)

surface is a migmatic overlying of a plutonic anatectic diapir. On the basis of the rock associations, ANASTASIU *et al.* (1979) divided the massif into two sectors: (I) Orotva-Putna Sector (northern part of the massif) and (II) Ditró valley-Güdüc-Békény Sector (central and southern part of the massif). Rocks of the Ditró valley-Güdüc-Békény Sector are divided into five rock complexes: (1) Complex of Ultramafic and Mafic Rocks; (2) Complex of Diorites; (3) Complex of Monzonites and Syenites; (4) Com-

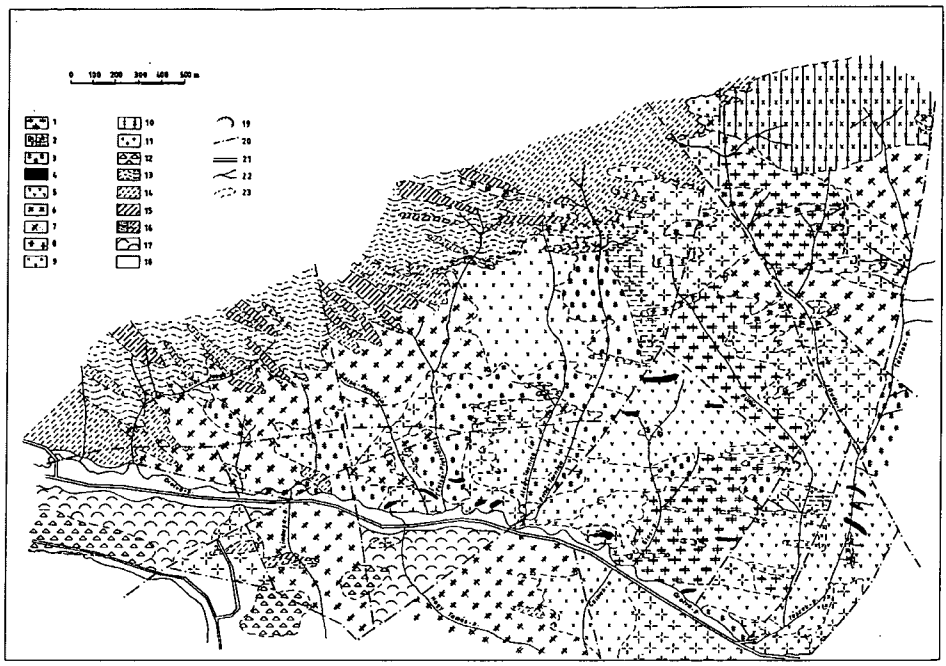

Fig. 2.
Geological map of the area of the northern part of the Ditró Syenite Massif between Csibi Jakab and Tászok creeks (PÁL MOLNÁR, ZÓLYA, 1988)
1. Meladiorite; 2a. Diorite without textural ordering; 2b. Diorite with textural ordering; 3. Hornblendite; 4. Biotitite; 5. Monzonite; 6. Alkali feldspar microsyenite; 8. Quartz syenite; 9. Syenite; 10. Nepheline syenite; 11. Granite; 12. Volcanic agglomerates; 13. Black quartzite; 14. Crystalline schists; 15. Mesometamorphites; 16. Hornfels; 17. Deluvial and coluvial sediments; 18. Alluvium; 19. Opened mine; 20. Fault; 21. Non-modernized road; 22. Creek; 23. Boundary of rocks

plex of Granitoids; (5) Complex of Foiditoidic Rocks. In their opinion, multiphase, multistage magmatic intrusions formed these rocks

Most researchers arranged hornblendites and meladiorites among the Complex of Ultramafic and Mafic rocks (ANASTASIU *et al.* 1979, ZÓLYA *et al.* 1985, ZÓLYA 1986; PÁL MOLNÁR 1988). On the basis of the IUGS standars (LE MAITRE 1989) only the hornblendites can be classified as ultramafic rocks (PÁL MOLNÁR 1992). Consequently, meladiorites should be petrografically arranged among the diorites.

Aim of this paper is tectonic and mineralogical-petrographical characterization of the diorite group.

PETROGRAPHICAL CHARACTERISTICS OF DIORITE GROUP

Diorites outcropping north of the Orotva creek and between the Csibi Jakab and Tászok creeks have a great tectonic variety. Hence, judgement of their petrography changed from time to time.

Diorites with ordering textures outcropping near the lower section of the Fülöp and Tászok creeks were arranged into essexites by IANOVICI (1933). In his opinion, texture of the rock represents schistose-fluidal transition. Approximately 70% of the

minerals are ordered, the color index (M) is about 50%. CODARCEA *et al.* (1957) suggest that diorites represent the dominant part of the Complex of Diorite-hornblendite Rocks. STRECKEISEN *et al.* (1974) emphasize that there are two dioritic complexes in the Ditró syenite massif: one is in the lower section of the Orotva valley, and another is in the zone of the Cengellér-Güdüc creek. Rocks of the last complex differ from the normal diorites in their essexitic chemical character. STRECKEISEN classified these rocks, together with monzodiorites, nepheline diorites and nepheline monzodiorites, as ditró-essexites. In opinion of ANASTASIU *et al.* (1979), dioritic rocks of Orotva from stratiform rock bodies, lens, schlierens, nests, and show great textural variety (pegmatoidic, normal and micrograined, ordered and non-ordered textures). The authors emphasize that the hornblendite-diorite transition is gradualy formed by the quantitative increase of plagioclase and the quantitative decrease of amphibole. On the basis of the color index (M), they described the following rock types: leucodiorites (M<25%), diorites (M=25-50%). ZINCENCO *et al.* (1978) suppose diorite complex of Orotva to be "ring" rocks because of the ordered texture of the most rocks.

On the basis of my field work and field experience, I suppose that diorites do not form stratiform bodies and lens. They are tectonically absolutely irregular bodies (*Fig. 2*), and show gradual transition and joggle toward the adjacent rocks. There cannot be found any definite spatial sequence (e.g., hornblendite → meladiorite → diorite → monzodiorete etc.) supposed by CODARCEA *et al.* (1957), and ring structure of various rock complexes suggested by ZINCENCO *et al.* (1978) cannot be showed, either. Shape of the diorite bodies is irregular, and their spatial arrangement is absolutely chaotic. Diorites can have a contact with any other rock type of the massif (PÁL MOLNÁR 1988), but generally occur near the hornblendites, surrounding them, and forming gradual transition or joggle. Direct contacts with the neighboring metamorphic rocks, however, are rare.

On the basis of the mineralogical composition, color index (M), and textural features, rocks of the dioritic group can be classified as follows:

- meladiorites with textural ordering
- meladiorites without textural ordering
- schistose meladiorites
- diorites with textural ordering
- diorites without textural ordering
- diorites with feldspar "schlieren"
- diorites with feldpsar aggregates
- leucodiorites with textural ordering
- leucodiorites withot textural ordering.

PETROGRAPHICAL DESCRITPION OF MELADIORITES

In general, meladiorites have close spatial connection with hornblendites. Beside the mineralogical composition, the textural difference of the rocks is the most characteristic. Most of the meladiorites has ordered texture. Reason of the ordering can be explained by several hypotheses: solidification of the rocks under high pressure (IANOVICI 1933), transformation of crystalline rocks by anatectic processes preserving the original schistose texture (CODARCEA *et al.* 1957), Fe-Mg metasomatosis of the crystalline rocks preserving the original schistose texture (JAKAB 1986). In general, the two rocks types have transitional contacts but direct contacts between meladiorites with ordered texture and hornblendites without ordered texture also occur.

1. Meladiorites with ordered texture

On the basis of the shape of the minerals, texture of the rock represents a transition between the hypidiomorphic and the allotriomorphic types. The ordered texture is due to the striped arrangement of the melanocratic minerals and the lathshaped plagioclase crystals (*Fig. 3.*).

Modal composition of the rock is shown by Table 1 and *Fig. 4.*

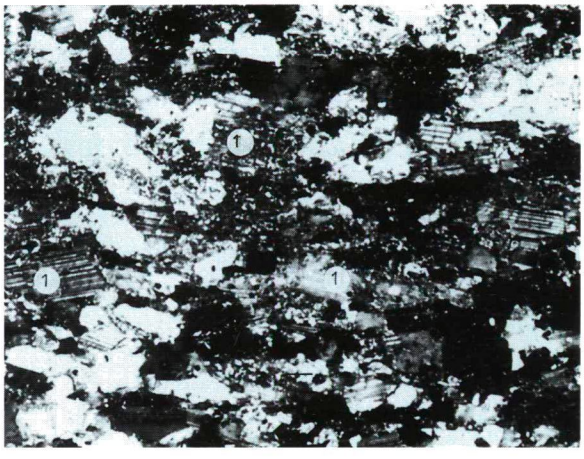

Fig. 3. Meladiorite with textural ordering
Striped arrangement of plagioclase crystals (x23; +N)

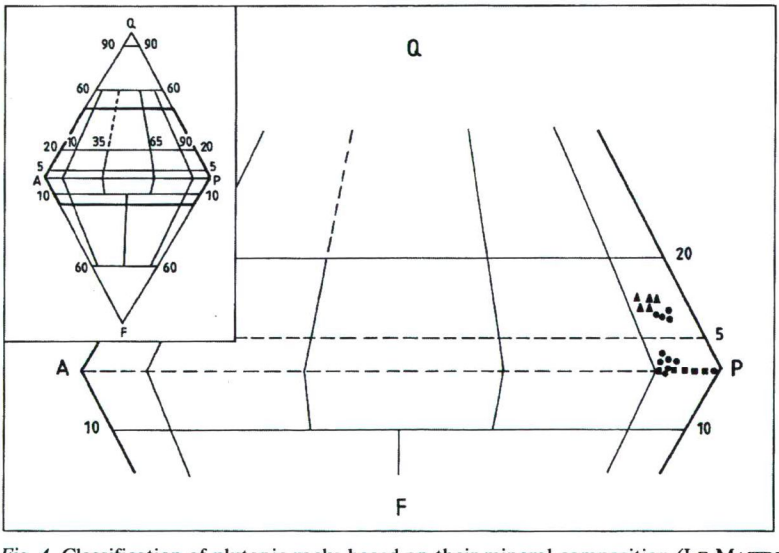

Fig. 4. Classification of plutonic rocks based on their mineral composition (LE MAITRE, 1989)
(●) meladiorites with textural ordering; (Δ) schistose meladiorites; (■) diorites with textural ordering

TABLE 1

Modal composition of the diorites

Minerals	Volume %			
	1	2	3	4
Aktinolite	0–13	–	–	–
Albite An_{0-5}	3–6	1–2	–	–
Apatite	2–6	4–6	4–5	–
Biotite	1–27	–	10–30	2–3
Carbonate	2–3	–	2–3	1–2
Chlorite	0–3	–	1–2	1–2
Hornblende	28–57	–	20–22	28–29
Plagioclase $An_{>5}$	9–44	14–18	38–60	54–55
Pyroxene	0–4	–	7–8	–
Orthoclase + Microcline	0–5	–	4–6	–
Epidote	0–3	–	3–6	–
Sericite	1–4	–	1–2	4–5
Quartz	4–6	2–3	–	5–6
Sphene	4–28	2–3	6–7	–
Magnetite	0–4	2–3	2–3	5–6
Chlorit + Secondary generation sphene + Calcite + Epidote + Biotite	–	70–80	–	–

(1) meladiorite with textural ordering, (2) schistose meladiorite, (3) diorite with textural ordering, (4) diorites with feldspar aggregates

Feldspars fill the space among the melanocratic minerals. Most of them are plagioclase, quantity of the alkali feldspars (microcline, orthoclase, perthite) is very low. Anorthite content of the plagioclase is 12–58% (CONSTANTINESCU and ANASTASIU 1977) but albite also occurs possibly as a secondary mineral. The feldspar crystals are very often sericitized, epizotized. Most of the feldspar crystals contain poikilitically idiomorphic apatite (*Fig. 5*).

Among the melanocratic minerals the long, prismatic hornblende is the most frequent one. It shows light brown-dark brownish green or light brown-dark brown pleochroism. Close intergrowth between hornblende and biotite can be found very often. The larger crystals are dented due to the actinolization. They contain poikilitically idiomorphic apatite and calcite (*Fig. 6*).

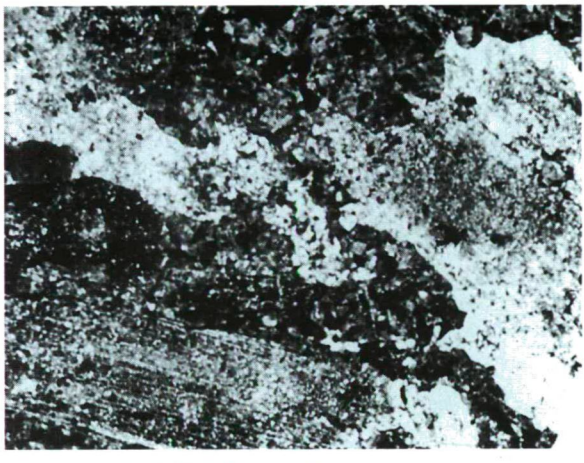

Fig. 5. Meladiorite with textural ordering
Sericitized, epidotized feldspar with apatite inclusion (x23; +N)

Fig. 6. Meladiorite with textural ordering
Hornblende(3) with apatite(1) and calcite(2) inclusions (x23; +N)

Very rarely, partly amphibolized pyroxene also occurs in the rock.

Biotite occurs as well-developed plates. The larger plates often have chlorite bands, and in some cases, are epidotized. Carbonate nests can be found among the plates, as well. Formation of the secondary sphene is also related to the larger biotite plates (*Fig. 7*).

The well-developed idiomorphic crystals of apatite can be found in all other minerals (even in the sphene crystals). Because of the rapid crystal growth, accumulation of the apatite crystals can be observed in the margins of the sphene ones.

The rare twinned sphene crystals are well-developed in every case (*Fig. 8*).

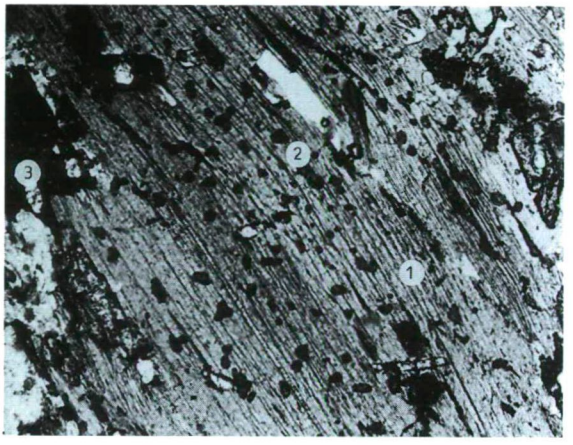

Fig. 7: Meladiorite with textural ordering
1. biotite; 2. secondary sphene; 3. apatite (x50; +N)

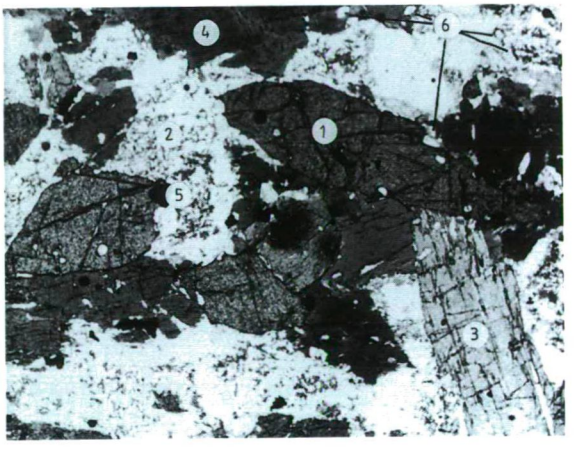

Fig. 8: Meladiorite with textural ordering
1. sphene; 2. feldspar; 3. biotite; 4. hornblende; 5. magnetite; 6. apatite (x23; 1N)

2. Meladiorites without textural ordering

Mineralogical composition of these rocks is the same as that of meladiorites with textural ordering. The only one difference is between their texture (*Fig. 9*). The amphibole and feldspar crystals have not oriented arrangement.

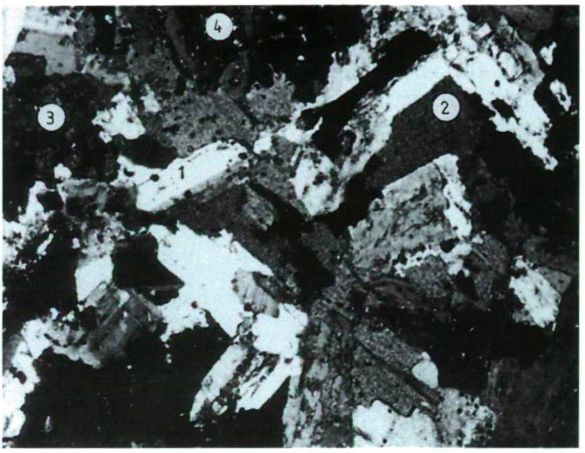

Fig. 9: Meladiorite with textural ordering
1. feldspar; 2. biotite; 3. sphene; 4. hornblende (x23; +N)

3. Schistose meladiorites

This rock type was first descibed as "laminated hornblenditic and gabbroid rocks" by CODARCEA *et al.* (1957). According to the authors, these rocks were formed by metamorphosis of hornblendites in the border zone of basic and ultrabasic rocks.

On the basis of the modal composition (Table 1), the rock should be named as meladiorite (*Fig. 4*). Microscopic texture of the rock is varying; on the basis of relative size of the minerals, it is porphyritic. The porphyritic texture is given by the rounded plagioclase crystals and the chlorite agglomerate containing xenomorphic megnetite inclusions. "Flow" appearance of this agglomerate provides the schistose character of the rock.

Melanocratic minerals (hornblende, biotite) are almost totally chloritized. Sphene, apatite and, rarely, nests and grains of quartz also occur. Parallel with the "laminating" planes, calcite grains can be observed.

Considering the frequent mechanical deformation of the minerals and massive appearance of the secondary constitutions (chlorite, carbonate), it is possible that these rocks are products of meladiorites that underwent tectonic movements. Appearance of calcite and quartz veins indicate non-metallifer hydrothermal solutions that circulated along the palaeofaults.

1. Diorites with textural ordering

Melanocratic minerals (mainly hornblende and/or biotite) form parallel bands. These parallel bands provide the ordered pattern of the texture (*Fig. 10*). Within the melanocratic bands, however, amphibole crystals are not parallel to the bands, and are arranged in divergent, radial or chaotic way.

Modal composition of the rock is ashown by Table 1. and *Fig. 4.*

Microscopic texture of the rock is hypidiomorphic. The most frequent leucocratic mineral is the plagioclase. The well-developed crystals are xenomorphic and trimmed (*Fig. 11*). The alkali feldspars (microcline, perthite) fill the space among the plagio-

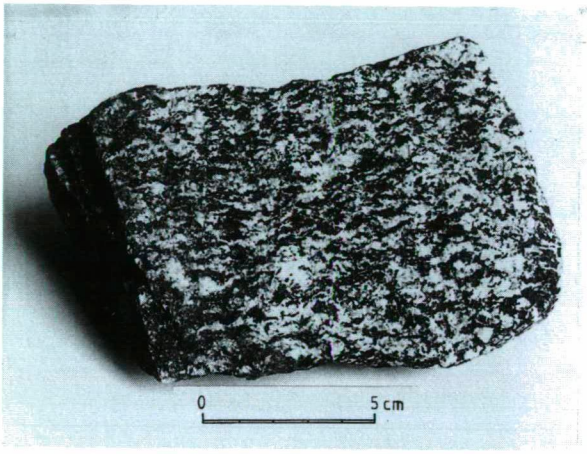

Fig. 10: Diorite with textural ordering

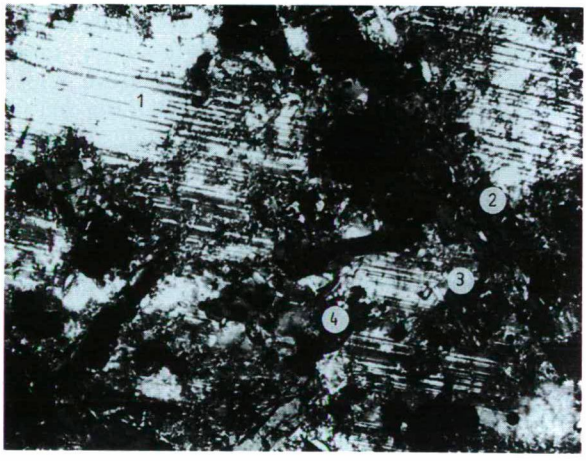

Fig. 11: Diorite with textural ordering
1. plagioclase; 2. biotite; 3. hornblende; 4. sphene (x23; +N)

clase crystals. Anorthite content of the plagioclase ranges from 10 to 45%, frequently 20–30% (JAKAB *et al.* 1987) that corresponds to oligoclase-andesine. It is frequent that cores of the crsytals are altered, sericite and apidote can be found as xenomorphic disperse grains. Chlorite veinlets can be observed in the fissures of the plagioclase crystals, as well.

Amphibole crystals of the rock are very similar to that of hornblendites (PÁL MOLNÁR 1992). The prismatic amphibole crystals have brownish-greenish pleochroism. They very often include sphene, rutile and apatite inclusions (*Fig. 12*). Rarely, the hornblende is chloritized, and in this way, the rock shows slightly greenish color. It is general that the hornblende is intergrown with the biotite.

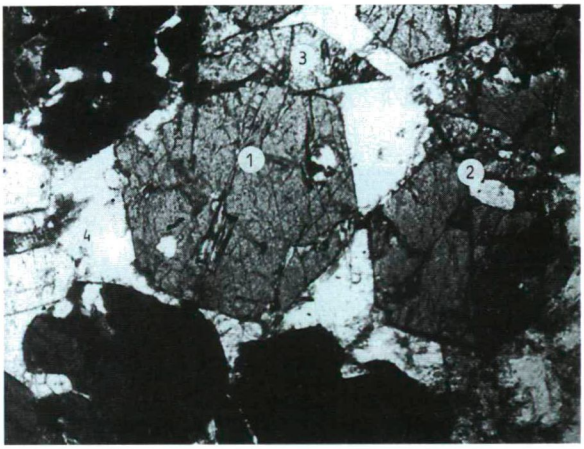

Fig. 12: Diorite with textural ordering
1. hornblende; 2. apatite; 3. sphene; 4. feldspar (x23; +N)

The prismatic, strongly fractured pyroxene crystals (diopside augite, Ti-augite, aegirine augite, augite) are quite rare in the rock.

The biotite has two variants. The well-developed lamellar one is chloritized on the edges of its crystals, has greenish-brownish pleochroism, and in some cases, has pyroxene desaggregate inclusions. The other variant occurs on the edges ot in cores of the amphibole crystals, and it frequently has apatite inclusions (*Fig. 13*). Epidotization also occurs on the marginal parts and along he cleavages of the biotite.

Accessory minerals of the rock are the same as that of hornblendites (PÁL MOLNÁR 1992). Sphene, apatite and opaque minerals (magnetite, titanomagnetite) are the accessory minerals of largest quantity (*Fig. 14*).

The most important secondary minerals formed by metamorphosis of the rock-forming ones are sericite, calcite, epidote and chlorite.

2. Diorite without textural ordering

Mineral composition of the these rocks is the same as that of diorites with textural ordering. The only one distinction criterion is the texture of the rock which is hypidiomorphic, phanerocrystalline, and has not textural oredring (*Fig. 15*). The grains-size ranges from 1.0 to 5.9 mm; in some cases, the rock is pegmatitic and microcrystalline.

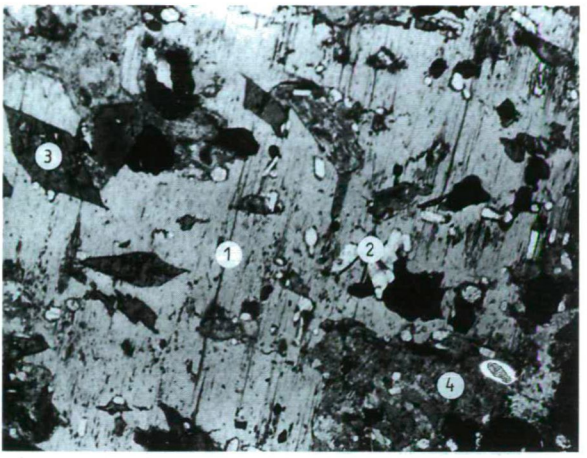

Fig. 13:
Diorite with textural ordering
1. biotite sericitized on the edges including apatite (2), sphene (3) and chlorite (4) inclusions (x50; 1N)

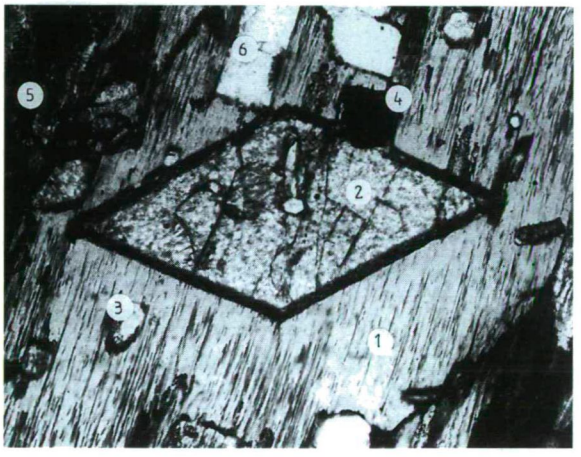

Fig. 14:
Diorite with textural ordering
1. biotite; 2. sphene; 3. apatite; 4. magnetite; 5. chlorite; 6. feldspar (x50; 1N)

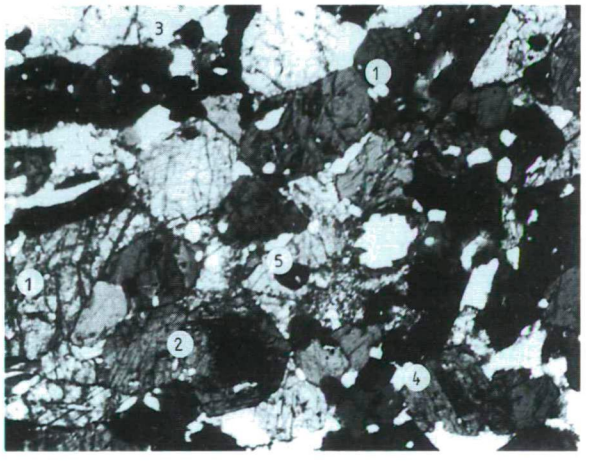

Fig. 15:
Diorite without textural ordering
1. sphene; 2. hornblende; 3. feldspar; 4. biotite; 5. magnetite (x23; 1N)

3. Diorites with feldspar "schlieren"

It is very frequent that the larger prismatic crystals of plagioclase are associated with the smaller isometric plagioclase ones. The smaller ones are weathered and, in general, occur in the melanocratic bands. Some smaller plagioclase crystals can be found in the larger ones. This observation suggests that the larger plagioclase crystals are the younger. Crystals of the plagioclase are parellel with together and with the melanocratic minerals, and in this way, determine the ordered arrangement of the texture.

4. Diorites with feldspar aggregates

There are plagioclase aggregates (cumulates) with diameter of 2–3 cm in the rock (*Fig. 16*). In these augens, hornblende of small size, small planes of biotite and, sometime, idiomorphic sphene crystals occur in form of continuous strings or in isolated grains. Melanocratic minerals of the feldspar augen are parallel to each other is every case. In most cases, their orientation corresponds to the textural ordering of the rock. It suggests that plagioclase crystals formed subsequently in such a way that they did not disturb the original orientation of the melanocratic minerals.

Fig. 16. Diorite with feldspar aggregates

Mineral composition of the feldspar schlieren and feldspar augen diorites is the same as that of diorites with textural ordering (Table 1).

By a decrease of quantity of the melanocratic minerals under 25%, diorites might gradually turn into leucodiorites. As in the case of diorites, leucodiorites with and without textural ordering can also be distinguished. Mineralogical parameters of leucodiorites are the same as that of diorites.

CONCLUSIONS

The above presented very detailed classification for rocks of the diorite group, which is based on their mineralogical compositions and textural characters, is the first one in the literature in such form.

From the geological map of the northern part of the Ditró syenite massif it emerges that diorites do not form layerlike or lenticular bodies. There are irregular rock bodies that have gradually transition and joggle toward the neighboring rocks. Diorites have always contact with hornblendites, and, beside rocks of the syenite group, they may have contact with metamorphic rocks, too. Consequently, diorites represent a connection of a certain kind between hornblendites and metamorphic rocks as well as between syenites, monzonites and granites. I suppose that two (spatially equivalent but temporally different) magmatic mass had to be intruded to form this tectonic setting. Because mineral composition of rocks of diorite group outcropping on the area between Csibi Jakab and Tászok creek and that of hornblendites, which are in contact with them, are very similar (PÁL MOLNÁR 1992), the mineralogical relationshp of these two groups is evident. It is possible that a subcrustal uktrabasic parental magma (hornblendites) was hybridized (creating the diorites) and taken toward the surface by a less siliceous megma. Naturally, the later magmatic intrusion, which is partly equivalent with the parental magma in space, assimilated the neighboring quartz-rich metamorphic rocks, and resulted in a series of acidic differentiated rocks.

Orientation of agglomerates (schlieren) of the maffic minerals might be a result of an intrusive brecciation, too. Thermal effect of the later magma might re-crystallize hornblende: the melt was forced into fissures of the hornblende, and surrounded the broken parts.

I emphasize that these petrogenetic conclusions are based on field and mineralogical-petrographical observations. Very accurate geochemical analyses (trace elements, isotopes) are need for confirmation or rejection of this model.

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